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Large-Scale Destabilization Events in Hydrological Structure of Oceans, Biotic Crises, and Corresponding Geological Records

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Abstract—Changes in the spatial organization of water masses in sea basins may represent one of the causes responsible for crises in the biosphere. Based on the analysis of experimental data and observations in modern sea basins, the ecological processes, which can be related to these events, are considered. Destabilization of seawater stratification can bring about high-productive ecosystems of the simplified structure. Changes in the natural equilibrium between the organic matter producers and consumers result in a significant excess of biological production, which, in turn, can lead to changes in sedimentation.

Key words: sea basins, water masses, hydrological structure, mass extinctions, detritus, anoxic events, sedimentation, biogeochemical cycle of carbon.

Changes in spatial organization of water masses in seas and oceans are thought to be responsible, along with other factors, for crises in the biosphere evolution (Wilde and Berry, 1984, 1986; Wilde *et al.*, 1990; Schindler, 1990; Nesov, 1995). These phenomena could stimulate substantial changes in biota, which, in fact, mark the stratigraphic boundaries (Sokolov, 1978). Biotic transformations are largely considered as resulting from sharp changes in abiotic environmental parameters (temperature, the content of oxygen dissolved in water, and others) substantially hindering or making impossible the further existence of some taxons. In addition, particular biotic factors, such as destruction of nutrient potential and others, are also believed to contribute to these transformations. Significantly less attention is paid to changes in the structural–functional ecosystem organization. Moreover, the analysis is, in fact, just a registration of more general tendencies like of the taxonomic diversity dynamics or changes in proportions of functional biotic groups. The fact is that any reconstruction based on paleontological and geological materials elucidates consequences of particular events, but not the events themselves. Crises in the biosphere history are reduced to two points: the relatively rapid extinction of some taxons and subsequent acceleration in radiation in some biotic groups due to the appearance of new ecological niches. It is evident that a more thorough study of ecological processes acting during the extinction periods of the old biota followed by appearance of the new one is an actual problem. Data provided by investigations in modern sea basins can essentially contribute to the available models.

The main objective of this work is summarize data on principal ecological processes resulting from destabilization of the water mass structure in modern sea basins. I consider observation results and experimental data obtained in the Black Sea during 1983–1990, as well as the data concerning the problem and available in publications. It is known that the ocean structure was essentially different during the "warm" and "cold" periods in the Earth's biosphere (Chumakov, 1995). From this standpoint, the modern Black Sea represents an extremely convenient object for investigation. Its upper, mobile oxygen-saturated zone is generally characteristic of sea basins of the modern "cold" biosphere, whereas the underlying, almost stable H_2S-contaminated zone makes it similar to past basins that existed under conditions of the "warm" biosphere (Wilde *et al.*, 1990; Nesov, 1995). Thus, the study of the sea can provide data useful for understanding processes characteristic of two different biosphere states.

MAIN TYPES OF THE BASIN STRUCTURE DESTABILIZATION

Before going to further consideration of the problem, I should specify some notions to be used. The water column of any large basin is heterogeneous in its physical–chemical and biotic properties and consists of separate water masses. Every water mass is formed in particular physical–geographic environments and is more or less stable in space and time (Dobrovol'skii, 1961). It is autonomous in surrounding waters and does not mix completely with the neighboring water mass. Such a behavior of water masses is mainly related to differences in their density that is, in turn, controlled by differences in the water temperature and salinity. The water column in modern (and past) sea basins is (was) stratified, i.e., it is (was) composed of autonomous

water masses arranged into several water layers. The neighboring water masses are (were) separated from each other by the pycnocline, a relatively thin layer characterized by a sharp change in the water density. The stable relative position of water masses is termed the hydrological structure of the basin (Beklemishev *et al.*, 1977).

Main water masses of seas and oceans exist during the long-term "geological" epochs (Beklemishev, 1969; Lisitsyn, 1980). In the course of evolution, most of marine organisms become adopted to dwelling in a particular water mass. It is clear that the destabilization of hydrological structure should unavoidably culminate in serious ecological consequences. Despite this, there is no clear definition of such phenomena. The main difficulty here is the fact that the hydrological structure experiences both the periodic (seasonal and long-term) and nonperiodic fluctuations, the latter mainly caused by changes in hydrometeorological conditions. This, however, does not lead to destruction of marine ecosystems in most cases; moreover, the fluctuations represent one of the attributes of their normal existence. In this connection, it is necessary to define which kind of changes should be considered as destabilization proper. Under this term, I understand changes that do not happen under normal conditions or go beyond the limits of the seasonal and long-term dynamics of the long-existing spatial hydrological structure. The destabilization brings about conditions atypical of the particular basin. Whereas usual fluctuations in the hydrological structure are the customary property of the basin, to which organisms adopt themselves in the course of evolution, the destabilizing events are extraordinary for them. It should be noted in this respect, that the more or less permanent (local) upwelling in oceans cannot be considered as destabilization, because it represents, in fact, one of the elements of the long-existing hydrological structure responsible for existence of many ecosystems (Vinogradov, 1977; Vinogradov and Lisitsyn, 1981). In this case, the upwelling cessation should be considered as a destabilizing event.

When describing phenomena related to destabilization of water mass stratification, the term "overturn" is often used. The latter, however, can be applied in its strict meaning only randomly, because destabilizations of the hydrological structure can be of different type.

Another very important aspect of the phenomena under consideration should be mentioned. In this context, only those significant changes in the water mass structure can be viewed as destabilizing, which occur during a relatively short time period. It can be assumed that, in the case of gradual, unidirectional, and very slow (not jump-like) environmental changes in the hydrological structure, representatives of the marine biota have enough time to evolve normally. It is apparent, that under these conditions, the ecological consequences will be principally different. Correspondingly, the regularities considered in the present work cannot be used for the analysis of situations like these (i.e., in reconstructions of the processes).

Causes, types, and scale of destabilizing events can vary. They can be related to different factors: volcanism, tectonic processes, transgression-regression cycles, meteorite impacts, and others events. Destabilization can be observed against the background of sea-level fluctuations and migrations of sea basin boundaries, or it may occur without substantial changes in basin configuration. Despite significant differences in the nature of these events, their ecological consequences can be similar. Considering the destabilization effects, one important point should be emphasized. The hydrological structure of both the modern (Lebedev *et al.*, 1974) and past (Lisitsyn, 1980) oceanic basins represents an integral system, and destruction of one of its elements can culminate in the avalanche-like chain of reactions and end in destruction of other elements. In this connection, the local or regional factor can provoke destabilizing events of a much larger scale, even of the global one.

Phenomena related to the destabilization of hydrological structure can be long-lasting, and the every phase of the process can be specific.

ENVIRONMENTAL CHANGES AT THE DESTABILIZATION STAGE

In addition to density, the deep water masses differ as a rule from surface waters in a significantly lower temperature and in some chemical parameters, among which most important for organisms are: (1) the dissolved oxygen content that commonly decreases with the depth; (2) the occurrence of toxic compounds (ions of heavy metals, H_2S) in deep layers; and (3) the downward decreasing concentration of nutrients (N, P, and others) needed for retaining biological productivity. The differences in the chemical composition of surface and deep-water masses to a large measure are mostly related to the permanent sinking of organic matter in the form of dead organisms or products of their life activity and to its decomposition at the bottom. The organic matter oxidation partly consumes oxygen dissolved in water. As a result, anoxic conditions can be formed, and this is usually accompanied by formation of ammonium compounds and H_2S (Wilde *et al.*, 1990). The end products of decomposition of sinking organic matter are represented, among other things, by inorganic nitrogen compounds and phosphates.

Two main types can be distinguished among various destabilizing events. One is mixing of two or more water masses. In this case, environmental conditions sharply change and, at the first moment, are controlled by physical–chemical properties of initial water masses and their proportions in the newly formed water mass. The second type of destabilization is the case, when the water mass either changes its spatial position, or becomes transformed, under the influence of various

external factors, into the water mass with new properties. For instance, when a deep water mass is raised to the surface, but not mixed with surface waters, the increased insolation and the contact with atmospheric air unavoidably change its main parameters. Similarly significant changes in water masses are characteristic for the case of their sinking to deeper levels. For the convenience sake, destabilization of the first type can be termed as *mictic* (from Greek "mixis," mixing), and another one as *choric* (from Greek "choros," the place). It is apparent that such a subdivision is very arbitrary. Any destabilization is accompanied by some mixing of water masses and, on the other, it can be almost always assumed that a part of the water mass remains beyond the contact zone during the mictic destabilization, although it changed its position. In other words, any type of destabilization can be defined as a principal trend of the event development. In case of the large-scale destabilization of hydrological structure, it would probably be useful to consider the mictic and choric zones separately depending on preponderant influence of the processes of one kind or another.

The influence of mictic destabilizing events on the marine biota primarily depends on the difference degree between physical–chemical properties of mixing water masses. It is clear that the greater the difference between parameters of water masses, the more significant should be negative consequences. In actual practice, it is difficult to a certain extent to define similarity and difference between water masses from the standpoint of their ecological significance. For instance, in case of the H_2S occurrence in one of the water masses, consequences of the mictic destabilization will be catastrophic for most representatives of the biota, even though all other parameters in the mixing waters are similar.

Changes in habitat conditions of planktonic organisms caused by destabilizing events of the choric type are generally characterized by smoother patterns, although they also occur during the relatively short time period. Unlike pelagic organisms that experience, during destabilizing events of the choric type, the influence of gradual, though fast environmental transformations in their habitual water mass, benthic organisms, whose life space corresponds to the particular bottom area, immediately fall into the other water mass with different physical–chemical parameters. It looks like they are "flooded" by the alien water mass.

When reconstructing conditions resulting from changes in the hydrological structure, it is common practice to evaluate significance of particular factors like the content of dissolved oxygen, different ions, and other components. When environmental changes are virtually unsuitable for most taxons, as it the case of the water rising from the H_2S-contaminated zone, such an approach can be quite productive. However, when environmental changes do not bear catastrophic consequences, but significantly change the biota, the approach would be inadequate sometimes. As is seen from experimental results and observations in modern basins, significant biotic changes can occur even in those cases, when environmental transformations are, at first glance, not catastrophic. For instance, waters from different layers of the oxygen-saturated zone in the Black Sea are devoid of highly toxic compounds, nonetheless organisms of different taxonomic groups suffer negative consequences from the alien water (Table 1).

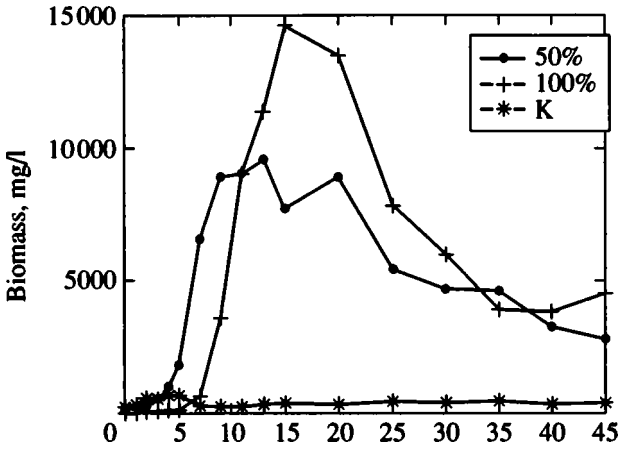
Table 1. Mortality of organisms (in %) dwelling in the sea surface layer after the hydrological destabilization of the mictic type (75% of the deep water mass)

Species	I	II
Planktonic Crustacea species		
Acartia clausi Giesbr.	0	16.5
Oithona minuta Kritcz.	11.5	18.5
Paracalanus parvus Cl.	11.0	20.0
Penilia avirostris Dana	61.5	100.0
Centropages kroyeri Giesbr.	36.5	38.5
Gastropods		
Bittium reticulatum (Costa)	6.3	9.0
Rissoa splendida Eichw.	0	1.7
Bivalves		
Mytilus galloprovincialis Lam.:		
planktonic larvae (veligers)	5.7	20.0
mollusks having shells 4 mm long	1.3	6.0
mollusks having shells 40 mm long	0	0
Mytilaster linaetus (Gmelin)	5.0	12.0
Pisces		
Fry of the Black Sea anchovy		
Engraulis enerasiholus ponticus Aleksandrov	21.7	73.3
Fry of the Black Sea horse mackerel		
Trachurus mediterraneus ponticus Aleev	20.0	81.7

Note: (I) water mass of the "intermediate cold layer" (depth of 60 m, 7.0°C); (II) water mass of the "subanaerobic layer" (depth of 120 m, 8.5°C).

DEVELOPMENT OF ORGANISMS IN AREAS WITH DISTORTED HYDROLOGICAL STRUCTURE

Destabilization of the hydrological structure leads to the formation of a new water mass with other physical–chemical parameters. Organisms, which survived the stress situation, fall into conditions often essentially differing from their habitual environments. Sometimes, these environments become more favorable for their development. For instance, the enrichment of surface waters with nitrogen and phosphorus compounds, or with other nutrients, which was caused by destabiliza-



The development of phytoplankton in experimental capacities filled with deep water (100%), with the mixed surface and deep waters (50%), and with the surface water (control experiment).

tion of the hydrological structure, may significantly increase productivity of some algae species. The experiment of adding waters from the lower layer of the anaerobic zone to the surface water results in the increase of the planktonic algae biomass by a factor of 2–3 and even more (Figure). However, this increase reflects the anomalous development of several opportunist species, whereas the abundance of other species decreases, or they disappear completely (Beznosov, 1995). Even deep waters of the H_2S-contaminated zone of the Black Sea highly toxic for most organisms, represent, after their short contact with the atmospheric air, excellent environs for some phytoplankton species (Polikarpov *et al.*, 1986, 1993). A similar biomass increase (Beznosov and Pobedinskii, 1988; Beznosov, 1989) was also observed in experiments with large bottom algae (macrophytes). The vegetable biomass increase can also trigger the development of some species consuming the latter. For instance, the artificial rise of deep waters (artificial upwelling) in the Black Sea induced both the increase of the macrophyte biomass and the abundance growth of small gastropods consuming macrophytes. Environments resulting from destabilization can also be favorable for development of some animal species, whose abundance was restricted in normal conditions by rivals or predators.

At first stages, the development of organisms depends to a significant extent on the destabilization type. The latter largely controls the rate and evolutionary trend of ecological processes. For instance, in the Black Sea experiments that imitated destabilizing events of the choric type, waters raising to the surface from lower layers of the aerobic zone were characterized by an insignificant productivity of phytoplankton because of its extreme scarcity there. In addition, phytoplankton consisted only of those algal species that dwelt in deep waters prior to destabilization, where their diversity was low and assemblage incidental to a significant measure. After destabilization of the mictic type, i.e., after mixing of deep waters rich in phosphorus and nitrogen compounds with surface waters, which contain physiologically full-fledged phytoplankton cells, the experiment resulted in the significantly higher productivity. In this case, the phytoplankton diversity became substantially higher. The development of bottom communities can also be controlled to a large extent by the destabilization type. In the case of choric destabilization, when benthic organisms are flooded by the alien water mass, their diversity in the subsequent period can be significantly lower as compared with that resulting from the mictic destabilization. It is apparent that, in case of episodic local destabilization of any type, the initial composition of both planktonic and benthic communities can rapidly be restored owing to migration of organisms from areas unaffected by destabilization. However, when the hydrological structure is more or less distorted throughout the entire isolated basin, or when this process (permanent or intermittent) is sufficiently long-lasting, the type of destabilization plays an essential role. Contrary to situation that is observed directly at the moment of destabilization, biotic factors induced by different types of interactions between organisms in the subsequent period can be important. For instance, experiments with the modern phytoplankton indicate that some species, which relatively well survived changes in abiotic conditions at the destabilization moment, disappeared later, when they did not withstand competition with other algal taxons (Beznosov, 1995). Almost always, experiments showed extinction of different zooplankton groups during the period of the intense phytoplankton development in the mixed deep and surface waters masses. The mass extinction of different zooplankton species in response to water poisoning with products of algae life activity during the periods of planktonic "blooms" ("red tides") was also repeatedly observed in natural environments of many basins, the Black Sea included (Koval', 1984). Thus, the disappearance of some floral and faunal groups from marine communities due to destabilization of the hydrological structure is of two categories. First, it is the death of organisms at the moment of destabilization of the hydrological structure in response to catastrophic environmental changes. Second, organisms die because of subsequent ecological processes.

CHANGES IN THE STRUCTURE AND FUNCTIONAL ORGANIZATION OF ECOSYSTEMS

To understand in sufficient detail the ecological processes peculiar of areas with the disturbed hydrological structure, it is necessary to analyze changes in the entire marine ecosystem, because the latter can significantly differ in their character from those that can be reconstructed using only the distribution patterns and abundance of individual taxons. Principal elements of eco-

systems are represented by: (1) producers of organic matter; (2) consumers of the matter; and (3) reducing organisms responsible for the complete mineralization of the matter. Almost all biological production of photosynthesis in marine basins is confined to the photic zone of the surface layer. Functioning of this part of the marine macroecosystem controls the development of biological communities in the deeper water layers as well (Vinogradov and Lisitsyn, 1981; Lisitsyn and Vinogradov, 1982; Lisitsyn, 1986). In this connection, the knowledge of possible consequences of the hydrological structure destabilization in the photic zone of sea basins seems to be very important.

Main producers in marine ecosystems are represented by phytoplankton species and, in the peritatal zone, by macrophytes (large bottom plants). The response of these organisms to environmental changes is ambiguous. On the one hand, consequences of environmental changes in response to destabilization (for instance, a sharp temperature fall) can be negative. On the other, the increased nutrient content (N, P, and others) in the surface layer can stimulate a significant growth of phytoplankton and macrophyte biomass (Toyota and Nakashima, 1987; Beznosov and Pobedinskii, 1988; Beznosov, 1995). Thus, despite the fact that many producers become less abundant or disappear from the biota under destabilization of the hydrological structure, the total productivity of the trophic level as a whole can increase. It should be emphasized, however, that the rise of nutrients from deeper layers to the surface is only the potential possibility of the biological productivity increase. In reality, biomass of phytoplankton increases only in the case, when the water mass formed due to destabilization of the hydrological structure remains stable over a long period sufficient for the algae development. When destabilization is short-term and unstable, its influence is reduced only to negative consequences of changes in dwelling conditions of marine organisms. In local upwelling zones, an increase in the phytoplankton abundance have not been observed, because the deep water mass is transported away from these zones by currents and become rapidly absorbed by surface waters. Nonetheless, these conditions can favor the intense development of attached algae (Beznosov, 1989). Contrary to planktonic forms, these algae can use nutrients from deep waters in the upwelling zone, even if the latter is as small as several hundreds of square meters.

Many animal consumers dwelling in the surface and deeper layers may perish because of different destabilization factors. Consequently, some species show the abundance reduction and decrease in the growth and reproduction rates (Beznosov, 1989, 1995). Thus, the life activity at this trophic level can be substantially reduced.

The fall of the surface water temperature that frequently accompanies the destabilization of the hydrological structure declines the reproduction rate of bacteria representing main reducing organisms (Beznosov and Pobedinskii, 1987). Nevertheless, the total abundance and biomass of bacteria in some experiments imitating similar situations sharply increased as compared with average values of these parameters in the surface layer (Beznosov, 1995). This is a consequence of the intense decomposition of organic matter accumulated in deeper waters due to their relative temperature rise related to upwelling.

Thus, destabilization of the hydrological structure substantially influence all trophic levels of the ecosystem in the photic zone, but the impact is principally different for every of them. Destabilization of the hydrological structure in the photic zone of a sea basin can produce a sharp growth of the vegetable biomass because of anomalous development of several opportunist phytoplankton and macrophyte species. Against the background of simultaneous suppression of consumers' life activity and reduction of their abundance, this unavoidably distorts balance between trophic levels, since a significant share of organic matter produced by photosynthetic organisms is not used at the higher levels of the trophic chain. The unbalanced production cycle like this, with disproportional phytoplankton development related to different factors, is distinguished in modern marine ecosystems of different oceanic areas (Cushing, 1959; Geinrikh, 1971).

DETRITUS FORMATION BY DESTABILIZATION OF HYDROLOGICAL STRUCTURE IN A BASIN

Comparing possible impacts of changes in the hydrological structure of sea basins on the ecosystem of the photic zone, one can conclude that these environmental transformations are favorable for appearance of "excessive" organic matter. This is organic matter produced by photosynthetic organisms, but not completely consumed within the surface layer and sunken to deeper sea layers. In modern sea basins with the stable hydrological structure, only 5–10% of the phytoplankton carbon is transported from the surface water mass downward (Izrael' and Tsyban', 1989; Kuznetsov, 1993; Tsyban' and Mosharov, 1995; Kuznetsov and Trotsyuk, 1995). In balanced ecosystems, organic matter precipitates in the form of pellets and small fecal clods representing the life activity product of filtering organisms, rather than in the form of the organomineral suspension consisting of plankton remains (Lisitsyn and Vinogradov, 1982; Lisitsyn, 1986). Finer fractions of organic suspension become largely decomposed in the water column never reaching the bottom. It is not inconceivable that some share of this fraction precipitates autonomously, but its proportion is virtually negligible in normal environments (Bogdanov *et al.*, 1971). However, under conditions of the distorted hydrological structure, this type of sedimentation may be preponderant. The available experimental data allow an assumption that the destabilization of hydrological structure may cause the considerably reduced activity

Table 2. Detritus content in experimental capacities at the terminal stage of experiments

Experimental conditions	Detritus content, mg of the wet weight/l
Destabilization of the mictic type (mesocosms containing mixed waters of the surface and "intermediate cold layer" in proportion 50 : 50, exposition in the surface layer)	2176.3 ± 692.1
Destabilization of the choric type (mesocosms containing water from the "intermediate cold layer", exposition in the surface layer)	1809.7 ± 559.4
Control experiment (mesocosms with water from the surface layer, exposition in the surface layer)	179.5 ± 64.8

of filtering zooplankton species. As a consequence, a significant share of organic matter is transported from the photic zone to deeper layers in the form of fine suspension. Similar phenomena can be observed in natural environments. In temperate latitudes, for instance, the spring bloom of phytoplankton slightly precedes that of zooplankton, which consumes the former. This situation leads to the unbalanced regime: a significant share of phytoplankton cells escapes the trophic chain sinking to deeper layers and enriching bottom sediments with organic matter and skeletal elements (Vinogradov and Lisitsyn, 1981). The sharp increase in detritus formation and change in the type of sedimentation manner also occur in the unbalanced ecosystems of the upwelling zones (Vinogradov, 1986), where the amount of organic matter leaving the photic zone can increase to 30–40% of the photosynthetic organic carbon (Gordeev *et al.*, 1983). It should be noted, however, that ecosystems of the upwelling zones are, as a rule, most unbalanced, when upwelling is active during the relatively short period (Geinrikh, 1971), i.e., when this process resembles the hydrological structure destabilization. When upwelling is permanent and always uniformly intense, the production cycle can be sufficiently balanced (Vinogradov and Shushkina, 1987), because the marine ecosystem in upwelling zones is adapted precisely to these environments. In addition, it should be remembered that the ecosystem evolution is always aimed at the most complete utilization of environmental resources (Gorshkov, 1985).

One should expect that in experiments imitating destabilization of the hydrological structure, the intense phytoplankton development in the absence of filtering zooplankton organisms is accompanied by formation of thick fine-grained suspension composed of dead algal and bacterial cells. Nonetheless, many experiments revealed the formation of abundant large (2–3 mm across) flakes consisting of phytoplankton cells and structure-lacking mass, but not of fine-grained suspension. Flocculation of the fine-grained organic suspension was apparently caused by abundance of surface-active substance assembling separate particles (Izrael' and Tsyban', 1989).

Table 2 demonstrates experimental results of the detritus formation in the 200-liter experimental capacities ("mesocosms") made of the transparent film. These mesocosms represented isolated marine environmental domains isolated together with planktonic organisms and exposed to insolation in the surface layer. The control experiments (with water from the surface layer) showed the relatively low amount of detritus formed and composed mainly of zooplankton pellets. In mesocosms with deep water, zooplankton partly perished at the initial stage of experiments because of changed environmental conditions and completely disappeared after the phytoplankton "bloom." It should be noted that the study of flakes under microscope revealed some amount of viable phytoplankton cells. Nonetheless, from the standpoint of the detritus formation, these cells can also be considered as the detritus constituents. Alive cells involved into the flake-formation process during flocculation seem to be able to exist in the surface layer for a short period. Available observations show that the flake buoyancy was *a fortiori* negative: sinking of most flakes after intense suspending by sampling occurred in less than 4–5 minutes. Alive algal cells in sinking flake-like aggregates of the "sea snow" were repeatedly noted in modern seas. For instance, in the Black Sea, these cells preserving their envelopes, nuclei, and pigmented protoplasm occur in the H_2S-contaminated zone (where their vital activity is impossible) at the depth of 300–500 m (Kondrat'eva and Serebrov, 1975). This undoubtedly indicates the substantially high precipitation rate of the cells.

In their work, Lebedeva *et al.* (1982) used for calculations of the sinking rate of dead zooplankton organisms the following equation: $V = 1.5\ l$ mm/sec, where l is the object length in millimeters.

My experimental study on the rate of flakes sinking showed the applicability of the formula for the considered case. As simple calculations show, the time of sinking of most particles beyond the photic zone averages from several hours to several days. It is clear that in the real situation, this process is unavoidably influenced by the water movements. Deceleration of particle sinking can be expected for the pycnocline zones (Izrael' and Tsyban', 1989), although under destabilization of the hydrological structure, the latter may disappear. Nonetheless, the unbalanced growth of the primary productivity in the surface layer should, in any case, significantly increase the detritus flux to deeper horizons of the water column. Sometimes, this may accelerate the organic matter accumulation in bottom sediments. This assumption accords with the fact that

bottom sediments richest in organic matter are accumulated in those areas of the World Ocean, where surface waters are habitats of unbalanced pelagic communities (Vinogradov and Lisitsyn, 1981). The increase in the organic matter influx into deeper horizons within sea areas, where the hydrological structure is disturbed, unavoidably changes the aspect of benthic communities even in those cases, when bottom layers have not been affected by the destabilization. Anoxic conditions may emerge in this case. The phenomenon apparently corresponds to the "eutrophic paradox" as Kuznetsov (1980) terms it. The essence of the paradox consists in the following: the unbalanced increase in the marine phytoplankton productivity not only precludes the higher productivity of bottom communities, but renders further existence of the latter completely impossible.

Thus, consequences of even local destabilization of the hydrological structure represent a multistage avalanche process, when environmental changes at the moment of destabilization just trigger the event. In particular situations, changes in the marine ecosystem can be variable, from the decreased abundance, diversity, and growth rates of separate species in marine communities to the complete degradation of the existed ecosystem, for instance, due to the mass extinction of taxons.

PROBABLE INFLUENCE OF THE HYDROLOGICAL DESTABILIZATION ON CLIMATE AND BIOCHEMICAL CYCLE OF CARBON

Inasmuch as the hydrological structure of sea basins represents one of the most important climate-forming factors, the large-scale destabilizing events unavoidably affect the climate to a significant extent. The effect depends on two different factors.

The deep-water masses commonly have the lower temperature, and destabilization of the hydrological structure frequently lowers the surface water temperature. This can bring about the local modification of the heat interchange in some areas of the hydrosphere–atmosphere boundary layer. Even more serious consequences can be induced by the functional changes in marine ecosystems. Since the hydrological destabilization is generally favorable for accumulation of biogenic sediments, one can assume that a significant amount of carbon can be transferred into sediments from atmosphere and hydrosphere. As known, the ratio between the total content of mobile carbon compounds and its amount enclosed in sedimentary rocks represents one of the most important factors controlling the global climate (Budyko and Izrael', 1987). It is likely the large-scale destabilization of the hydrological structure of the ocean that substantially and periodically influenced the carbon cycle in past geological periods (Malinovskii, 1993). Global climatic changes in response to these events culminated in principal transformations of the biosphere. Concluding, I should attract attention to another point. Even single short-term, but large-scale destabilization of the hydrological structure, which does not trigger the long-lasting changes in oceanic circulation and sea-level fluctuations, enriching only the photic zone with nutrients for a short period, can provoke, nevertheless, the irreversible climatic changes. Therefore, it is difficult to agree with Nesov (1995), who stated that the large meteorite impacts could not affect climatic conditions of the globe and the structure of global ecosystems.

INDICATIONS OF LARGE-SCALE HYDROLOGICAL DESTABILIZATION IN GEOLOGICAL RECORDS

The processes induced by destabilization of the hydrological structure in seas and oceans are highly variable. It can be assumed that their results are frequently reflected in geological and paleontological records as autonomous events heterogeneous at first glance. Moreover, phenomena related to the same event can be separated in time. Consequently, it is necessary to define particular evidences of the hydrological destabilization, which could be preserved in the geological record.

(1) Direct evidences of the hydrological destabilization.

(1.1) Destabilization of the hydrological structure is frequently accompanied by the mass death of marine organisms. This event, often selective with respect to individual taxons, is controlled by particular conditions and largely depends on the destabilization type. In the case of destabilization of the choric type, plankton usually suffers to a grater extent than benthic organisms "flooded" by the alien water mass. As compared to event of the choric type, environmental changes by the mictic destabilization are less harmful for benthic organisms than for plankton. However, the drastic environmental changes are very brief in terms of the geological time scale, and, sometimes, they cannot be recognized in the paleontological record. Environmental transformations at the moment of destabilization are usually distributed irregularly in space, and the former biota thus can be locally preserved during a certain period. In this case, a rapid restoration of biota based on preserved organisms is quite probable. This happens even after the rise of highly toxic H_2S-contaminated waters, since the contact of these waters with the atmosphere leads to their detoxication (Polikarpov *et al.*, 1986), and corresponding zones in the surface layer become suitable for life and can be populated again.

It should be also noted that the onset of hydrological destabilization is not always accompanied by the extinction of organisms. For instance, destabilization of water stratification can eliminate the anoxic conditions in bottom layers, and organisms can populate the latter.

(1.2) Destabilization can be manifested by sharp variations in chemical (isotopic) composition of bottom

sediments and in composition of fossils. For instance, a large-scale destabilization of the hydrological structure in oceans at the Permian–Triassic boundary is evident from the data on the isotopic composition of fossils (Hoffman, 1989). In the opinion of Hoffman, the terminal Permian was a period of the geologically instant influx of oxygen-rich waters to the bottom layers. On the other hand, this period coincides with a sharp decrease of the dissolved oxygen content in the surface layer.

(1.3) It is possible to assume as well that the large-scale destabilization of the hydrological structure can be accompanied by activity of strong bottom currents and, correspondingly, by erosion and redeposition of sediments.

(1.4) An "instant" (from the geological standpoint) change in the bathymetric distribution of benthic biocoenoses can be an evidence of the distorted hydrological structure. Phenomena like this are recorded in modern sea basins. For instance, upwelling is accompanied by the upward migration of deep water biocoenoses at a shallower depth level as compared with their other habitats (Ghelardi and North, 1958).

(2) Possible manifestations of changes in the structural–functional organization of ecosystems in response to destabilization of the hydrological structure.

(2.1) The large-scale destabilization of the hydrological structure can result in the mass extinction events that were induced in different periods by secondary processes (for instance, by intoxication of organisms during the phytoplankton "bloom" and mass death).

(2.2) Environmental changes caused by destabilization of the hydrological structure may result not only in the mass death of organisms, but also can provoke an intense development of some taxons. Therefore, the destabilization events may be marked by a sharp increase in the abundance of opportunist species.

(2.3) When related to the unbalanced growth of bioproductivity, the destruction of the structural–functional organization of ecosystems may represent a long-lasting multistage process. It takes some time to eliminate the old and to create the new stable ecosystem that completely utilizes environmental resources. The ecosystem evolution principally differs from evolution of individual species. The new ecosystems appear in the course of selection of most fortunate combinations of different constituents: producers and consumers of organic matter (Zavarzin, 1993). It can be assumed that the fairly long period after destruction of the former water masses corresponds to a succession of ecosystems able to exist during a relatively short (in geological sense) time and characterized by a low diversity of biota and significantly simplified trophic chain. Since organisms need a long time interval sufficient for their evolution and formation of new forms more adapted to new environments, the short-lived ecosystems should mainly consist of species inherited from the earlier ecosystem, although the species role can be essentially changed. The accompanying decrease in abundance and diversity of species and reduction of their distribution areas imitate patterns of "slow" extinction in the paleontological record.

(2.4) Destabilization of the hydrological structure can be manifested as changes in composition of sedimentary material and sedimentation rates. The following observation can probably be useful for paleontologists. Under normal environmental conditions, the fine-grained organic remains reach bottom sediments, as was mentioned, in the form of fecal pellets. It should be taken into consideration, however, that the organic suspension is subjected during sedimentation to mechanical and biochemical reworking, being recurrently consumed by zooplankton. Under distorted hydrological conditions, this factor can be negligible. In addition, flake-like aggregates of the "sea snow" can quite rapidly transport living planktonic organisms as well. In this connection, one can expect that the preservation of microfossils could be better in periods following the events of hydrological destabilization.

(2.5) Destabilization can result in appearance of anoxic environments in response to unbalanced production of organic matter.

(3) Climatic changes related to destabilization of biogeochemical processes are unavoidable. Such phenomena influence the biosphere as a whole and are always marked in the geological record. Nonetheless, events of the hydrological destabilization are separated from their consequences by a significant time gap. Therefore, it is sufficiently difficult to establish the relationship between the events and its final result.

Data considered in the paper lead to the following two conclusions. First, the most noticeable large-scale events destabilizing the hydrological structure of seas and oceans can be separated in time from their results that are registered in the geological and paleontological records (for instance, changes in lithology, extinction of taxons, etc.).

Second, the extent of biotic changes related to events of the hydrological destabilization is eventually controlled by subsequent changes in the bioproductivity balance within marine ecosystems, but not by the toxicity of deep waters rising to the surface. When deep aerated waters are sufficiently rich in main nutrients, their rise to the photic zone can be as catastrophic as the rise of H_2S-contaminated waters.

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Phanerozoic Striate Miospores as Indicators of Biotic Events and Stages of Paleoflora Evolution

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Abstract—The development history of striate miospores is considered. These miospores appeared first by the end of the Silurian–the beginning of the Devonian. During the existence period of striate miospores that lasted more than 400 m.y., the number of "striate genera" varied and was different in every geological epoch of the Middle–Late Phanerozoic. The variations in abundance of "striate genera" distinctly depended on and, probably, were controlled by major phases of orogenesis, transgressions, regressions, glaciations, aridity fluctuations, and humidification of climate. These oscillations can be used for recognition of homeostasis periods, critical events, and transitional stages in the Phanerozoic evolution flora, the most characteristic of which are briefly described. Plants with striate miospores apparently filled the ecological niches vacated after extinction of their predecessors and, therefore, the striate miospores can be used as indicators of biotic events and stages in the paleoflora evolution. Permian paleoflora from southwestern Caspian depression is considered as a particular example elucidating such events and stages. Two important biotic events are recorded in this region in the boundary intervals separating the lower and upper Artinskian substages and the Ufimian and Kazanian stages. These events divide the Permian flora history into three periods corresponding to the Lower, Middle and Upper series of the Permian System.

Key words: striate miospores, taxa, Phanerozoic, Permian, biotic events, stages, paleoflora, palynoflora.

INTRODUCTION

In practice of studying the development of biosphere of the Earth, the recognition of biotic events is particularly important, because they mark boundaries between the essential changes in climate, composition and organization of biota, and also in sedimentation regime. In order to identify these cardinal events correctly, it is necessary to know their indicators, which would allow us to determine transformations of this kind and to understand other phenomena they are connected with and caused by. According to my opinion based on an extensive practical experience in the study of the Middle–Late Phanerozoic palynofloras, the striate miospores can be regarded as the universal indicators of flora evolutionary stages and biotic events.

The striate miospores represent a group of palynomorphs with sculptural elements like threads, ribs, strands, furrows, and striae of predominant unidirectional orientation (Kremp, 1967). These miospores are easily recognizable owing to the oriented arrangement of the sculptural elements. Accordingly, they invariably attract attention of researchers being easily recognizable in a mass of dispersed miospores recovered from the Middle–Upper Phanerozoic deposits of different ages. This is evident from photographic materials presented in numerous palynological publications.

While studying the Phanerozoic palynoflora, I noted the presence of striate sculptural elements in many species of ferny, gymnosperm, and angiosperm plants. Moreover, the initial many researches did not classify these forms as striate. In addition, despite the great number of publications in this country and abroad, where regularities in stratigraphic distribution, morphology and classification of variable miospore species are considered, there are no papers dedicated to the general development history of striate miospores. Therefore, my efforts were concentrated on establishing and counting the genera incorporating species with striate sculpture in order to determine their significance in evolution of Phanerozoic floras and to outline importance of striate miospores for diagnosis of phytostratigraphic stages and biotic events.

The present paper describes the results of this research. However, before the discussion of factual materials, it should be noted that the estimated abundance rates of striate genera couldn't be considered as ultimate. With the further growth of palynological information, the number of striate genera may become greater. For example, as compared to my data published earlier (Podgainaya, 1993; 1995; 1996a; 1996b), the number of striate genera considered in this paper increased by factor of 1.5–2. Oddly enough, the development trends of striate miospores revealed initially remain unchanged.

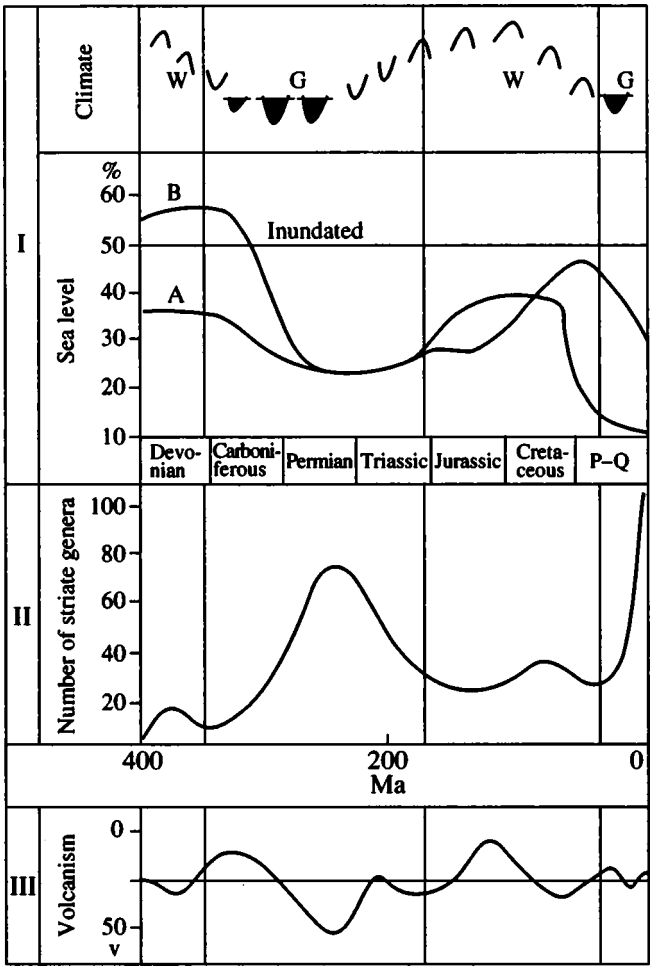


Fig. 1. Secular variations in abundance of striate miospore genera versus changes in the geological–geographical environs: sea-level oscillations (A) after Vail *et al.* (1977) and (B) after Hallam (1977), both curves are quoted from work by Fisher (1986); variation curve of volcanic activity after Brooks (1952) and intensity volcanism (v, m/Ma) after Veklich (1987) with modifications; (W) warming; (G) glaciation, filled sockets indicate time ranges of ice-sheet glaciations; (P–Q) Paleogene–Quaternary.

STRIATE MIOSPORES AND STAGES OF PHANEROZOIC FLORAS EVOLUTION

The first striate miospores (*Emphanisporites*) appeared at the Silurian–Devonian boundary simultaneously with origin of terrestrial higher plants. During more than 400-m.y.-long history of striate miospores, the amount of their genera varied and was different in separate geological periods. For instance, 18 form-genera characterize the Devonian, 36 the Carboniferous, 72 the Permian, 45 the Triassic, 22 the Jurassic, 37 the Cretaceous, 20 the Paleogene, and 104 the Neogene–Quaternary. My own observations (Podgainaya, 1995) and comparisons with other data on paleoclimatology (Schwartzbach, 1955; Sinitsyn, 1980; Veklich, 1987) show that these variations in amount of striate genera depend on climatic changes and peculiarities of lithogenesis and tectogenesis in certain geological periods. These variations distinctly follow the main phases of such geological processes as the orogenesis, sea transgressions and regressions, glaciations, fluctuations in aridity and climate humidification, and sedimentation.

The revealed tendency in historical development of striate miospores is graphically shown in a form of a curve depicting secular variations in their abundance rate (Fig. 1). The curve correlated with the plot of Phanerozoic supercycles (Fischer, 1986) and with the graph showing the time-dependent dynamics of volcanism (Veklich, 1987). Fig. 1 clearly demonstrates that the phenomenon of striate miospores was most characteristic of the Permian and the Neogene–Quaternary and coincided with the greatest sea regressions and periods of enhanced orogenic and volcanic activity, and also with peak phases of ice-sheet glaciations. Moreover, according to Fig. 1, an increase in abundance of striate taxa during the Cretaceous was, probably, caused by a combination of certain specific climatic and tectonic conditions.

The epochs of preparation for cardinal reconstructions, the periods of the ecosystem crises and homeostasis are marked by completely different abundance rates of striate taxa in the vegetal communities. For example, the minimal amounts of striate taxa coincide with epochs of transgressions, humid climate, and humid-type lithogenesis. Floras of these epochs show an extensive territorial and occasionally global distribution of some taxa, the reduction of phytogeographical differentiation, and dynamic stability in paleoflora composition. Stages in flora evolution during these epochs can be termed the homeostasis stages. Floras of the Late Devonian–initial Early Carboniferous, Late Triassic–Early Jurassic, and terminal Late Cretaceous–Paleocene periods exemplify these stages, and precisely in these palynofloras, the number of striate taxa was minimal.

The reformation stages of vegetal communities took place after the long-term preparatory stages, when abundance of striate taxa either grew (Middle–Late Carboniferous and beginning of the Early Permian, Middle–Late Jurassic, terminal Paleogene–initial Neogene), or became reduced (terminal Late Permian–Middle Triassic, the Late Cretaceous). Among striate taxa appeared during these preparatory stages, there were certain highly specialized species with a rather short life span. These were, for example, *Columinisporites hejleri* Doubinger, *Fusacolpites permicus* Djupina, *Vittatina zaueri* Efremova, and others. Actually these species, as also some others, are indicators of biotic events.

The peak abundances of striate miospores coincide with the greatest ecosystem crises caused by the orogenic processes, sea regressions, stages of development of thick ice sheets, arid climatic phases, and intensification of arid lithogenesis. By analogy, the stages of flora evolution typical of these epochs are called critical.

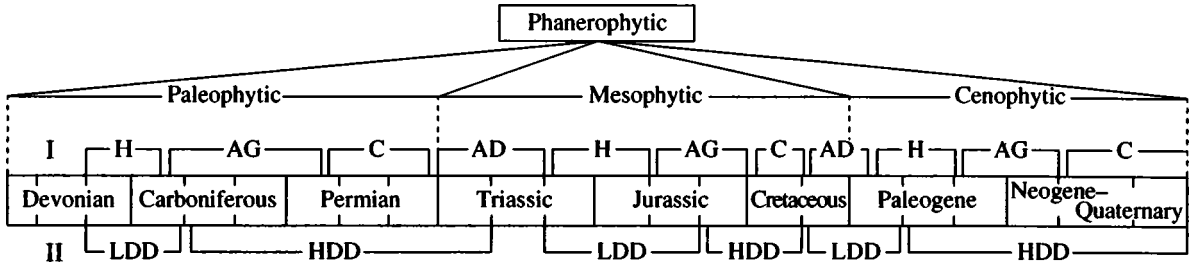


Fig. 2. Stages in evolution of Phanerphytic floras (compiled by N. N. Podgainaya): (I) development stages of striate miospores, such as homeostasis (H), abundance growth (AG), crisis (C), and abundance decline (AD); (II) stages after Vakhrameev *et al.*, (1970, p. 35) based on plant remains; (LDD) low differentiation degree; (HDD) high differentiation degree.

Floras of this kind were characteristic of the terminal Early Neogene, the larger part of the Late Permian, the Early Cretaceous, and the Middle Neogene–Quaternary. The distinct climatic and phytogeographic differentiation typical of these stages and caused by the aforementioned abiotic factors was associated with more intense extinction of certain taxa, some highly specialized striate miospores included. Gradually, some combinations of casual factors resulted in the evolutionary outbursts reflecting the very rapid diversification of plant species, in their adaptive radiation and extensive expansion. As a consequence, these events caused tremendous reformations in vegetal communities, like the transition from the Carboniferous hygrophytic flora to the Permian xerophytic gymnosperms, and from the Mesozoic mesophilic coniferous–filicine flora to flora of the Cenozoic angiosperms.

Figure 2-I showing the succession of paleoflora stages is plotted with due account for the phenomenon of striate miospores and their development trends. This diagram is well correlative with the diagram of paleoflora evolutionary stages suggested by Vakhrameev *et al.* (1970) (Fig. 2-II) and based on changes in climatic and botanical–geographical environments. However, the diagram reflecting peculiarities in development of striate miospores, which also depends on changes in physical–geographical conditions, better and more circumstantially details events in the vegetal kingdom.

Apparently, the striate miospores carried out the historical mission of pioneer taxa occupying the ecological niches vacated after extinction of their predecessors. This is clearly evident from the plots illustrating the extinction dynamics of the Permian palynoflora in the southwestern part of the Caspian depression and coeval changes in abundance of striate miospores (Fig. 3). The part of curve B corresponding to the greatest abundance of striate taxa fills the gap between the two peaks of the extinction curve A. Plants with striate miospores were probably evolutionary plastic, bioproductive, and sufficiently adaptable to migration. Evidently, they produced a tremendous amount of miospores with an extraordinary stable sporoderm, which well protected the reproductive cells during their life time in very adverse, sometimes extreme climatic conditions, and also during the subsequent fossilization and burial under unfavorable conditions of the arid lithogenesis. That is why the finds of striate miospores are frequent and can be used as indicators of certain biotic events of variable character and significance.

STRIATE MIOSPORES, BIOTIC EVENTS, AND STAGES IN THE PERMIAN PALYNOFLORA EVOLUTION

The history of palynoflora from Permian deposits in the southwestern part of the Caspian depression is discussed here as an example illustrating how striate miospores mark the biotic events and stages in flora evolution during the respective period of time. My choice is based on the following reasons: first, the Permian System corresponds to the time of prevalence of gymnosperms with striate miospores; second, a considerable amount of data on the composition of the Permian palynoflora is at my disposal now. The information is obtained on the basis of more than three thou-

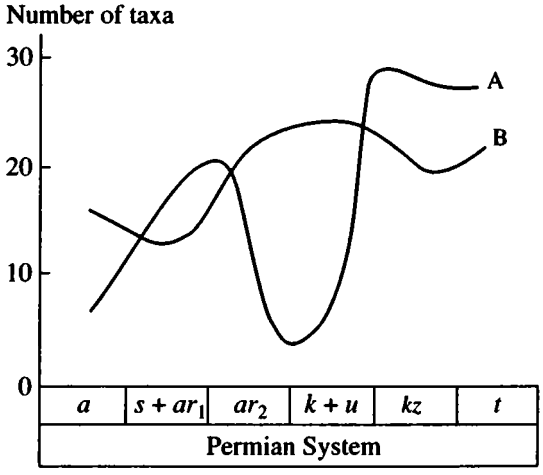


Fig. 3. Comparative dynamics of Permian palynoflora extinction and abundance variations of striate miospore taxa: (A) extinction curve (after Podgainaya, 1994); (B) secular variations in abundance of striate miospores (species); (a) Asselian; (s) Sakmarian; (ar) Artinskian; (k) Kungurian; (u) Ufimian; (kz) Kazanian; (t) Tatarian.

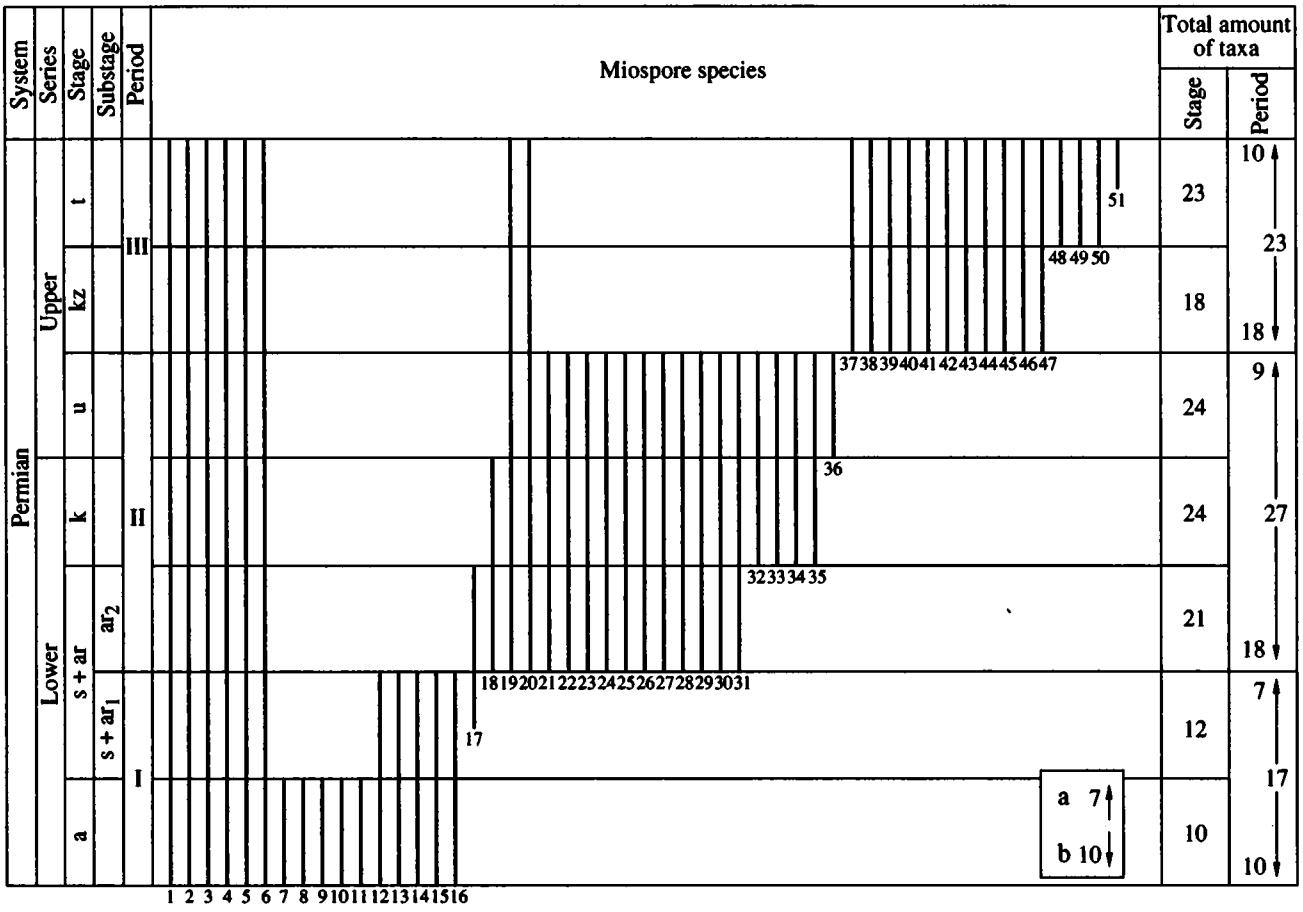


Fig. 4. Scheme of striate miospores distribution in Permian deposits of the southwestern Caspian depression (compiled by N.N. Podgainaya): (a) number of taxa transiting in overlying deposits; (b) number of extinct taxa within the unit range (stage indices as in Fig. 3). Miospore species: 1, *Vittatina striata*; 2, *V. vittifer*; 3, *V. costabilis*; 4, *Hamiapollenites bullaeformis*; 5, *H. tractiferinus*; 6, *Protohaploxypinus latissimus*; 7, *Hamiapollenites exilis*; 8, *H. saccatus*; 9, *Complexisporites fistulatus*; 10, *C. polymorphus*; 11, *Columinispora heilery*; 12, *Striatodiplopinites permica*; 13, *S. lucidus*; 14, *Vittatina zaueri*; 15, *V.* ex gr. *scutata*; 16, *V. vittifer* f. *foveolata*; 17, *Fusacolpites permicus*; 18, *V. subsaccata*; 19, *Protohaploxypinus perfectus*; 20, *Striatoabietites* spp.; 21, *Vittatina persecta* f. *angusticostata*; 22, *V. crassa*; 23, *V. vittifer* var. *cribrata*; 24, *V. persecta*; 25, *Protohaploxypinus verus*; 26, *P. vulgaris*; 27, *Striatodiplopinites parviextensisaccus*; 28, *S. silvestritipus*; 29, *Protohaploxypinus micros*; 30, *Striatodiplopinites parvisaccus*; 31, *Striatopodocarpites tojmensis*; 32, *Protohaploxypinus suchonensis*; 33, *P. chaloneri*; 34, *Striatoabietites elongatus*; 35, *Vittatina cinctutus*; 36, *Paucistriatopinites* sp.; 37, *Striomonosaccites* spp.; 38, *Taeniaesporites* sp.; 39, *T. noviaulensis*; 40, *T. ortisei*; 41, *T. labdacus*; 42, *Lueckisporites virkkiae*; 43, *L. microgranulatus*; 44, *Striatopodocarpites ellipticus*; 45, *S. cancellatus*; 46, *S. antiquus*; 47, *Corisaccites* sp.; 48, *Vittatina verrucata*; 49, *V. ovalis*; 50, *Crustaesporites* sp.; 51, *Gnetaceaepollenites* sp.

sand samples recovered from core sections of deep boreholes drilled in the West Caspian depression during the geological prospecting for oil and gas.

In contrast to large phytochronological stages of the Phanerozoic, which are distinguished using the secular variations in abundance of striate genera, the diagnosis of biotic events and stages of flora evolution during the chosen period is based on the analyzed changes in abundance of striate miospore species. The quantitative taxonomic and morphostructural analysis of palynofloras, in combination with data on lithological–facies peculiarities of the host rocks, revealed several biotic events unequal in significance and rank. It was established that the process of renewal in palynoflora composition was irregular and stepwise. However, two significant events distinctly divide the evolutionary history of Permian palynoflora into three approximately equal time periods (Fig. 4).

The first Asselian–early Artinskian period was characterized by a high abundance of the gymnosperms and Filicinae, both representing relics of the Late Carboniferous flora. Typical of this substage is the *Potonieisporites–Gardenasporites–Vattatina–Sporae* assemblage of miospore. The indicated succession of taxa reflects their quantitative proportions in palynofloras from the mentioned and somewhat lower levels. The miospore assemblages of the time period in question are macerated from sediments of predominantly transgressive megacycle represented by carbonate, carbonate-terrigenous, flyschoid and flyschoid-molassic beds,

which were deposited under relatively favorable climatic conditions and predate the salt accumulation period. That is why the striate taxa from these beds are not numerous. Only 17 of 49 most frequent miospore species from the Asselian–lower Artinskian deposits are striate (Fig. 1, Podgainaya, 1994). A group of persistent striate taxa originated at that time includes *Vittatina striata* Luber, *V. vittifer* Luber, *V. costabilis* Wilson, *Hamiapollenites bullaeformis* (Samoil.) Jans., *H. tractiferinus* (Samoil). Jans., and *Protohaploxypinus latissimus* (Samoil.) Hart.

The xeromorphic pollen of these species experienced different adaptive morphostructural changes (Podgainaya, 1988) and controlled for many centuries the general aspect of Permian palynofloras. Morphological peculiarities of striate miospores of the first stage are vaguely defined. Most frequent amid them are forms with scarce and wide (5 μm and wider) parallel ribs (Plate 1; Figs. 9, 10, 11). Forms with numerous ribs wedging out are rare. At the end of the stage, some pollen species grew in size (Plate I; Figs. 1, 12).

Changes in the physical–geographical conditions and the commencement of renewal in the vegetal groups are recorded in the Sakmarian–early Artinskian interval. At that time, the geographic area under consideration experienced orogenesis that was followed by growing aridity of climate. A sudden change in species and generic composition of plant assemblages occurred at the boundary between the lower and upper Artinskian substages. The palynoflora responded to abiotic events by reducing the diversity of *Potonieisporites*, *Gardenasporites*, and *Sporae* genera and by parallel diversification of xeromorphic *Striatiti Disaccites* and *Vittatina* pollen. By that time, 29 of 49 taxa of the Asselian–early Artinskian assemblage became extinct. Among them, there were almost all filicine species, the relics of the Late Carboniferous flora, and also 10 striate taxa inherited in the Asselian strata from the Carboniferous: *Hamiapollenites exilis* Koltsch., *H. saccatus* Wils., *Complexisporites polymorphus* Jizba, *C. fistulatus* Krus., *Striatodiplopinites permica* Krus., *S. lucidus* Krus., *Vittatina zaueri* Efr., *Columinisporites heyleri* Doub., and others. In total, only seven of 17 striate taxa of the first period survived it and continued their evolution further. Six of these seven taxa are the persistent species mentioned above (Fig.4).

The second late Artinskian–Ufimian period of the Permian flora evolution corresponded to the formation time of typical Permian flora of xerophytic gymnosperms. This was the epoch of active morphostructural diversification and extensive radiation of new species, including the ribbed miospore forms (Figs. 3, 4). Palynoflora characteristic of this stage is represented by the *Striatiti Disaccites–Vittatina–Cordaitina–Entylissa* assemblage of miospores from the transgressive–regressive megacycle of salt-bearing sediments. The studied region was at this time under influence of further orogenic processes and recurrent short-term transgressions and regressions. The piedmont troughs were filled here with gray molassic deposits, whereas the vast bays and lagoons accumulated thick beds of evaporites. The plant kingdom reacted to these events by the relatively rapid changes in phytocoenoses, as it is evident from by the seven successive palynological assemblages (Levina and Podgainaya, 1989), and by the active diversification in the *Striatiti Disaccites, Vittatina*, and *Coraitina* genera.

More than 30 new plant taxa appeared in this period, and 28 of them had striate miospores (Fig. 4). Two new taxa (*Protohaploxypinus perfectus* Naum. and *Striatoabietes* spp.) enlarged the group of persistent species with striate miospores. Disturbances in habitat environments resulted in appearance of extremely large and even monstrous miospores. Morphology of striate miospores of the second period was diverse. At the beginning of the stage, there were slightly costate forms (Plate II: fig. 8, *Striatodiplopinus parvisaccus* Efr. and fig. 12, *Protohaploxypinus samoilovichii* Hart). Afterward, forms showing numerous narrow wedging ribs with perpendicular branches became dominant (Plate II: fig. 4, *Protohaploxypinus vulgaris* Efr. and fig. 19, *Vittatina persecta* Sauer). Forms with very thickened sporoderm, like that of *Vittatina crassa*

Plate I. Miospores of the Asselian–early Artinskian stage of paleoflora evolution:

(1) *Potonieisporites* sp., Borehole Sukhota 1, depth interval of 2767–2773 m, Sample 12, prep. 1, lower Artinskian Substage; (2) *Potonieisporites lemniscatus* Schwartsman var. *uralicus* Koltschina, Borehole Dzhakuevo 4, depth interval of 2808–2814 m, Sample 911, prep. 72, Asselian Stage; (3) *Gardenasporites oberrauchi* Klaus, Borehole Sukhota 1, depth interval of 2767–2773 m, Sample 12, prep. 1, lower Artinskian Substage; (4) *Gardenasporites pinnatus* Krusina, Borehole Astrakhan 26, depth interval of 3887–3893 m, Sample 13, prep. 5252/ 1, Asselian Stage; (5) *Pseudopodocarpus expresus* Krusina, Borehole Dzhakuevo 4, depth interval of 2808–2814 m, Sample 911, prep. 72, Asselian Stage; (6) *Complexisporites fistulatus* Krusina, Borehole Sukhota 1, depth interval of 3498–3501 m, Sample 5, prep. 175, Asselian Stage; (7) *Florinites luberae* Samoilovich, ditto; (8) *Limitisporites* sp., Borehole Vostochnyi Sharnut 4, depth interval of 2834–246 m, Sample 121, prep. 7672, Asselian Stage; (9) *Hamiapollenites bullaeformis* (Samoilovich) Jansonius, Borehole Dzhakuevo 4, depth interval of 2808–2814 m, Sample 911, prep. 72, Asselian Stage; (10) *Striatodiplopinites permica* Krusina, Borehole Astrakhan 26, depth interval of 3887–3893 m, Sample 13, prep. 5252/1, Asselian Stage; (11) *Vittatina costabilis* Wilson, ditto; (12) *Vittatina zaueri* Efremova, Borehole Dzhakuevo 4, depth interval of 2808–2814 m, Sample 911, prep. 72, Asselian Stage; (13) *Vittatina vittifer* Luber, Borehole Sukhota 1, interval of 2767–2773 m, Sample 12, prep. 1, lower Artinskian Substage; (14) *Vittatina foeolata* Tschudy et Kosanke, Borehole Dzhakuevo 4, depth interval of 2808–2814 m, Sample 911, prep. 72, Asselian Stage; (15) *Cyclogranisporites punctulatus* (Waltz) Luber, Borehole Sukhota 1, depth interval of 3498–3501 m, Sample 5, prep. 175, Asselian Stage; (16) *Lycospora pusilla* (Jbrahim) Schopf et Bentall, ditto.

Magnification for all species 600; all species from collection of N.N. Podgainaya.

Plate I

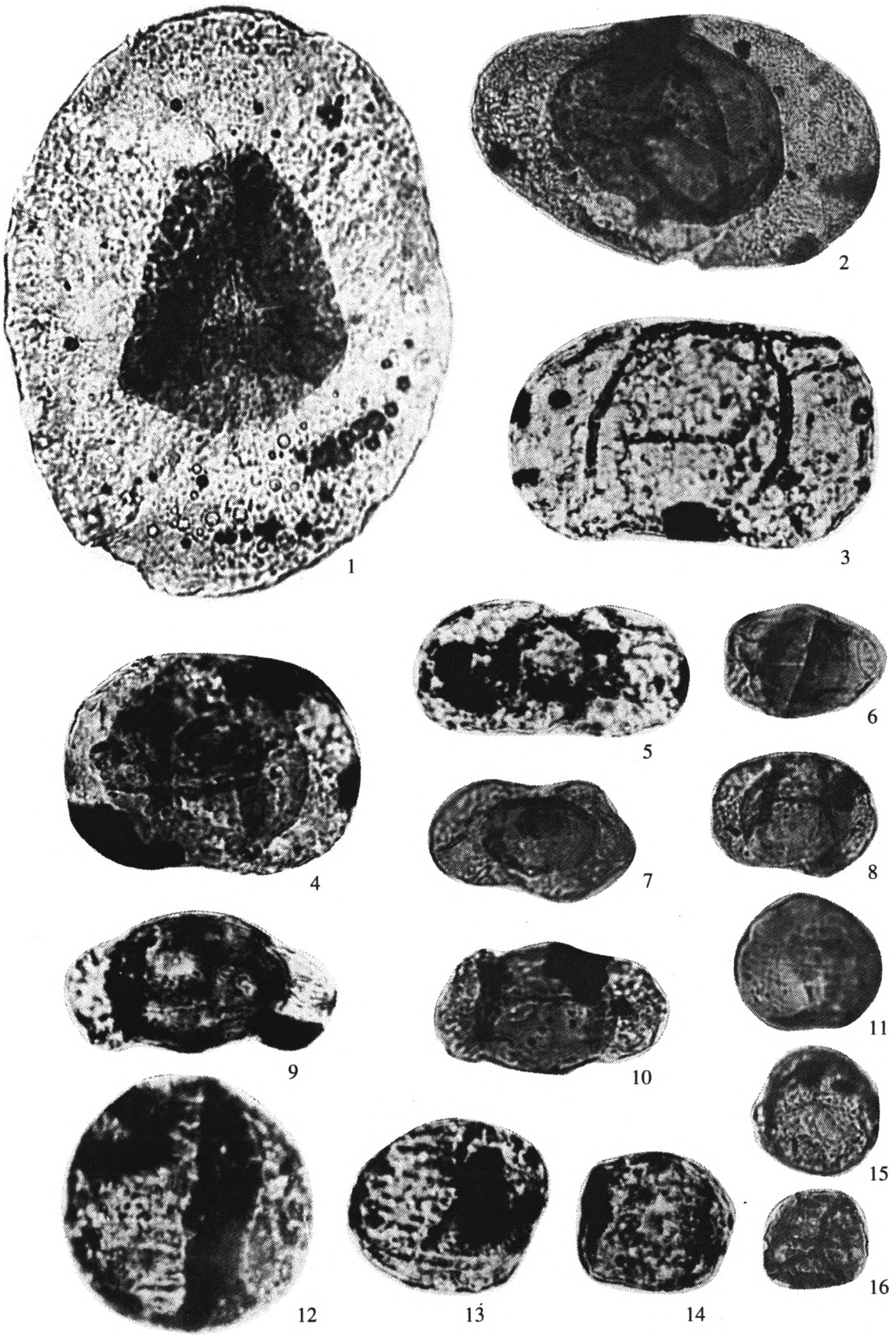
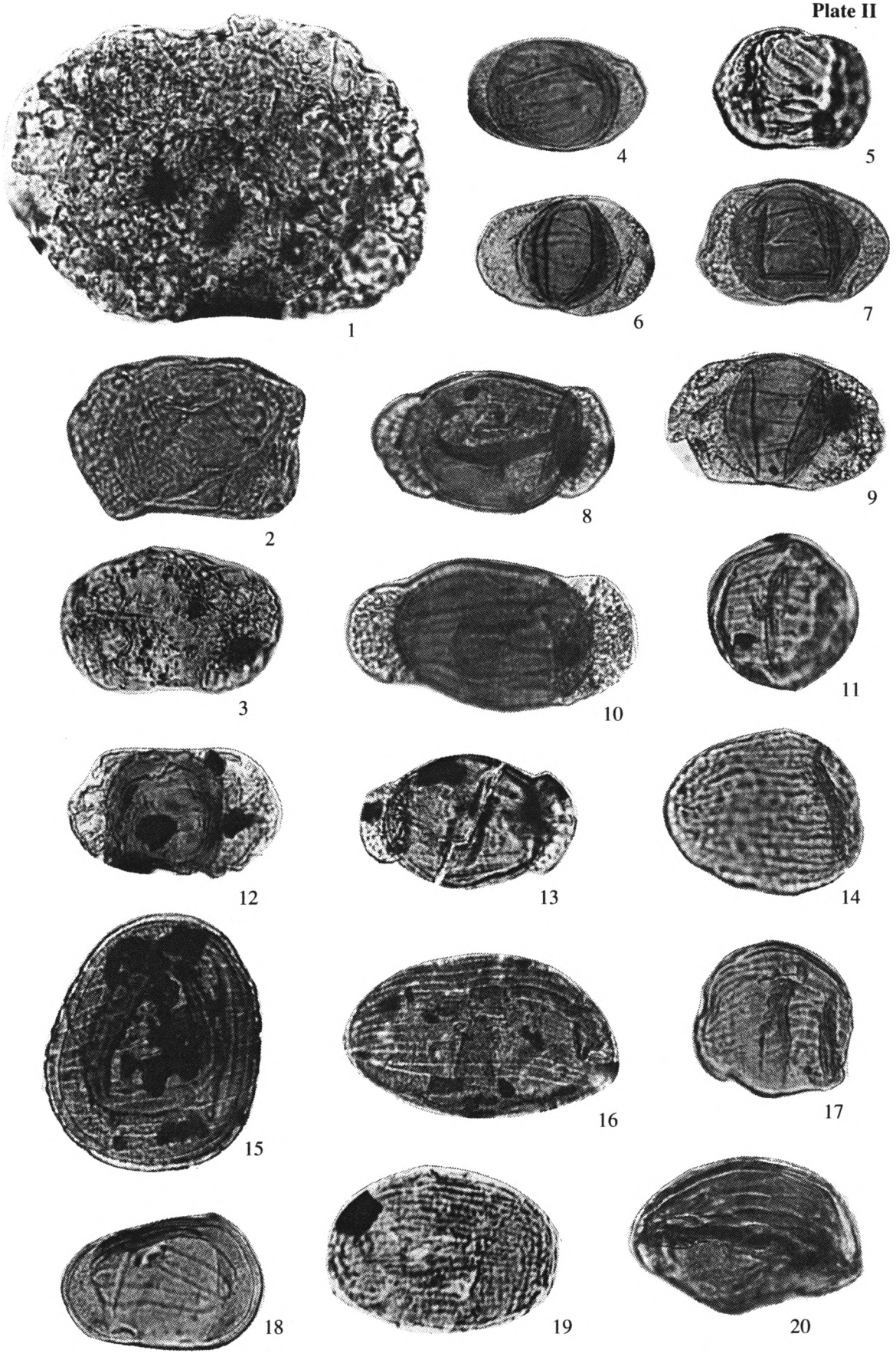


Plate II



Sauer (Sauer, 1965; Plate VI, fig. 3), appeared at the end of the stage.

The most sudden changes in palynoflora occurred at the boundary between the Ufimian and Kazanian stages. The growing aridity of climate in this boundary time and closure of the saliferous basin resulted in transition to predominantly continental type of sedimentation. During the time period in question, 37 of 50 plant taxa typical of the upper Artinskian–Ufimian deposits became extinct, and 18 of them had striate miospores. Only 10 taxa with miospores of this kind survived this event, eight of the persistent type included (Fig. 4).

The third Kazanian–Tatarian period in the flora evolution is marked by the initial transition from the Permian xeromorphic flora to a new flora of a more mesophytic aspect. The *Striomonosaccites–Lueckisporites–Taeniaesporites–Vittatina–Sporae* miospore assemblage of that time is associated with a thick regressive sequence of sandy-silty deposits postdating the salt-bearing megacycle. In this assemblage, 21 of 36 most typical miospores are striate. Thus plants producing these miospores were still dominant, but the speciation rate is lowered. At the level under consideration, only 11 new taxa with striate miospores appeared and became typical. Most of these miospores are of a new morphological structure; these are either grains with 4–6 taeniae (*Taeniasporites*, Plate III; fig. 13), or with two proximal corymbs divided by a diametrical leasurae (*Lueckisporites*, Plate III; figs. 11, 12), and small ellipsoid forms with numerous narrow ribs decorating their dense dark-colored body (*Striomonosaccites*, Plate III; fig. 3). The general aspect of palynofloras was controlled by presence of *Striomonosaccites* Bharadnaj f. *speciosus* Warjuh., *S.* cf. *varius* Sauer, *Taeniaesporites ortisei* Kl., *T. labdacus* Kl., *T. noviaulensis* Lesch., *Leuckisporites vikkiae* Pot. et Kl., *L. microgranulatus* Kl., and *Gnetaceaepollenites* sp. The appearance of these species obviously foreshadowed the beginning of development of new coniferous–filicine flora of the Mesozoic time. This is evident from appearance, at the end of the third stage, of filiciform spores of *Nevesisporites* sp., *Densoisporites nejburgii* (Schulz) Balme, *Discisporites colliculiniformis* (K.-M.) Warjuh., and others, which became typical plants of the Triassic.

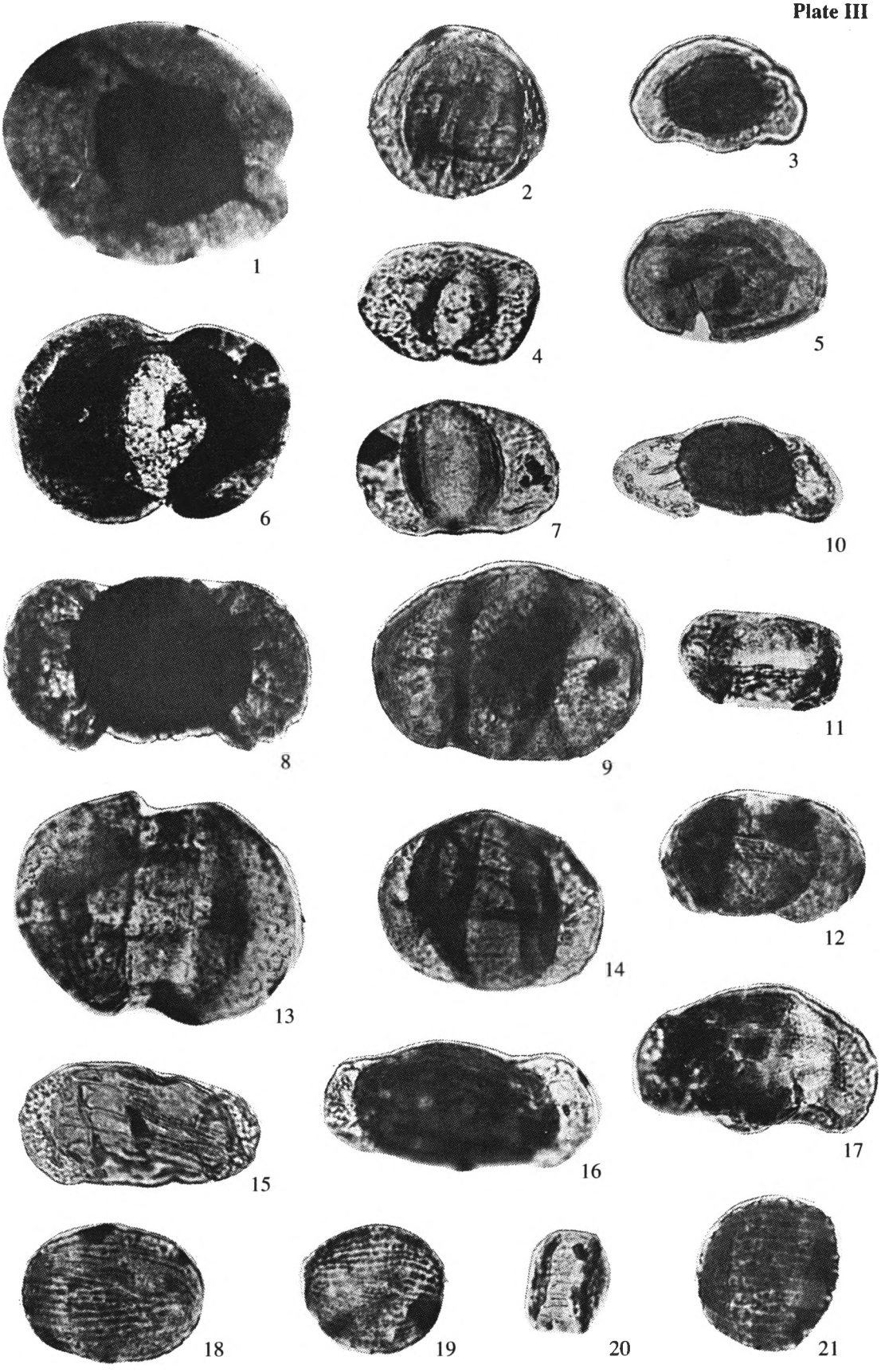
The presented analysis of compositional changes in the Permian palynofloras clearly shows that there were two biotic events, one at the boundary between lower and upper Artinskian substages and another at the boundary between the Ufimian and Kazanian stages (see Figs. 3, 4). These events divided the history of the Permian flora evolution into three characteristic stages: the Asselian–early Artinskian, the late Artinskian–Ufimian, and the Kazanian–Tatarian periods. Paleoflora of each stage is characterized by its own taxonomic and morphological set of miospore assemblages: one predating the salt-bearing sedimentary megacycle, the second characterizing the salt accumulation period, and the third postdating the latter (Plates I–III). Thus, the Asselian–early Artinskian and the Kazanian–Tatarian palynological assemblages are transitional. The Asselian–early Artinskian *Potonieisporites–Gardenasporites–Vittatina–Sporae* assemblage is predecessor of xerophytic flora of the late Artinskian–Ufimian period, which was extremely rich in plants producing striate miospores. This flora appeared as a consequence of the considerable ecosystem crisis that was in progress during the entire second stage. The Kazanian–Tatarian time was a preparatory period of floral renewals, which gave rise to the Mesophytic Mesozoic flora.

The general evolutionary trend is inferable from the distribution dynamics of striate miospores near the conventional boundaries of biotic events. For instance, if the first period, when 10 taxa of striate miospores became extinct and 25 new taxa originated between the lower and upper Artinskian substages, represents the flourishing time of xerophytic gymnosperm flora, then the second one between the Ufimian and Kazanian substages, when 18 their taxa perished leaving place for ten

Plate II. Miospores of the late Artinskian–Ufimian stage of paleoflora evolution:

(1) *Florinites* sp., Borehole Karasal 5, depth interval of 3839–3852 m, Sample 13, prep. 2330/2, upper Artinskian Substage; (2) *Protohaploxypinus* sp., Borehole Karasal 5, depth interval of 3737–3743 m, Sample 5, prep. 3038/1, Kungurian Stage; (3) *Protohaploxypinus* sp., Borehole Karasal 5, depth interval of 4080–4087 m, Sample 13, prep. 3086/1, upper Artinskian Substage; (4) *Protohaploxypinus vulgaris* Efremova, Borehole Listinka 1, depth interval of 1794–1797 m, Sample 14, prep. 4586/1, Kungurian Stage; (5) *Striatites angulistriatus* Klaus, ditto; (6) *Protohaploxypinus perfectus* (Naumova) Sauer, Borehole Severnye Shadry 399, depth interval of 967–973 m, Sample 76, prep. 5634/1, Ufimian Stage; (7) *Protohaploxypinus verus* Efremova, Borehole Listinka 1, depth interval of 1794–1797 m, Sample 14, prep. 4586/1, Kungurian Stage; (8) *Striatodiplopites parvisaccus* Efremova, Borehole Shiryaevo 18, depth interval of 3806–3812 m, Sample 51, prep. 5421/1, Kungurian Stage; (9) *Protohaploxypinus varius* (Bharadwaj) Balme, Borehole Severnye Shadry 399, depth interval of 967–973 m, Sample 76, prep. 5634/1, Ufimian Stage; (10) *Hamiopollenites tractiferinus* (Samoilovich) Jansonius, ditto; (11) *Vittatina costabilis* Wilson, Borehole Karasal 5, depth interval of 3920–3931 m, Sample 20, prep. 2334/2, upper Artinskian Substage; (12) *Protohaploxypinus samoilovichii* Hart, ditto; (13) *Hamiapollenites bullaeformis* (Samoilovich) Jansonius, Borehole Severnye Shadry 399, depth interval of 967–973 m, Sample 76, prep. 5634/1, Ufimian Stage; (14) *Vittatina flavata* (Maheshwari et Kar) Stanichnikova, Borehole Shiryaevo 14, depth interval of 3438–3444 m, Sample 5, 4110/1, Kungurian Stage; (15) *Vittatina* sp., ditto; (16) *Vittatina simplex* Jansonius, ditto; (17) *Vittatina costabilis* Wilson, Borehole Severnye Shadry 399, depth interval of 967–973 m, Sample 76, prep. 5634/1, Ufimian Stage; (18) *Vittatina striata* Luber, Borehole Shiryaevo 14, depth interval of 3438–3444 m, Sample 3, prep. 4110/1, Kungurian Stage; (19) *Vittatina persecta* Sauer, Borehole Shiryaevo 14, depth interval of 3806–3812 m, Sample 51, prep. 5421/2, Kungurian Stage; (20) *Vittatina vittifer* Luber, Borehole Severnye Shadry 399, depth interval of 967–973 m, Sample 76, prep. 5632/1, Ufimian Stage.

Plate III



newly appeared, can be considered as a prelude to the Permian xerophytic flora decline.

All the data discussed above show that the division of Permian System into three periods not only reflects the natural stages in the paleoflora evolution, but also elucidates the most important periods of historical–geological development of the region.

The materials on the Permian flora history discussed here are of prime importance for the current discussion concerning the revision of the general stratigraphic standard of the Permian System. The new variants of the Permian chronostratigraphic scale (Leven *et al.*, 1997) seem to be based on the incomplete information, because they take into consideration exclusively the trends of marine fauna evolution. As I think, it is impossible to get the high-resolution scale for any period of the event stratigraphy, especially for the geocratic periods of the Middle–Late Phanerozoic, if we ignore the biotic events on continents. Therefore, when revising any stratigraphic scales and schemes, it is necessary to consider the palynological data as well. Miospores of terrestrial vegetation are actually those euryfacies fossils, which are very important for the general correlation of marine and continental deposits. According to experience of researchers in this country and abroad, "palynochrons" (the term introduced by E.D. Zaklinskaya in 1985) are traceable over the great distances. Moreover, specialists in palynology discovered a series of indicators, which help to interpret the meaning of many physical–geographical events. Striate miospores are among such indicators. It is also important that the palynological data well agree with the available information on the evolutionary history of other organisms. For example, two important floral crises in the mid-Artinskian time and at the Ufimian–Kazanian boundary probably triggered, to some extent, two global crises in the evolution of Permian tetrapods "at the end of the Artinskian and during the terminal Guadeloupian" (cited from report by S. Lucas, United States, in Leven *et al.*, 1997, p. 11). In addition, the Artinskian crisis that affected the terrestrial flora and fauna was almost concurrent with that recorded in marine sedimentary sections (Leven *et al.*, 1996). It is obvious that the Artinskian events and those that occurred at the Ufimian–Kazanian boundary should be taken into account when determining the boundaries between subdivisions of the Permian System.

CONCLUSION

The considered information on the history of development of striate miospores implies that they are important fossils for diagnosis of particular stages in paleoflora evolution and of biotic events that divided the contiguous stages. The analyzed dynamics of striate miospores distribution in ancient palynofloras indicates that they are reliable indicators of changes in the physical–geographical environments and of successive stages in the paleoflora evolution. The striate miospores reveal not only the levels of biotic events, but also the character of environmental changes in proximity to these events. Abundance rates of striate miospores elucidate trends of the evolutionary and ecosystem processes typical of certain geological epochs. The successive stages of paleoflora evolution, if they are based on the phenomenon of striate miospores, reflect events in the plant kingdom better than any other schemes, elucidating their particular details.

Thus, this study shows that the striate miospores may serve as an important instrument of phyto- and palynostratigraphy, as the means of obtaining the valuable information of scientific–theoretical and practical significance.

ACKNOWLEDGMENTS

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Plate III. Miospores of the Kazanian–Tatarian stages of paleoflora evolution:

(1) *Monosaccites* sp., Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 6, prep. 2522/1, Kazanian Stage; (2) *Striomonosaccites* Bharadwaj, ditto; (3) *Stromonosaccites* sp., Borehole Ivanovka 3, depth interval of 2605–2613 m, Sample 3845, prep. 2815/2, Kazanian Stage; (4) *Vesicaspora* ex gr. *magnalis* (Andreeva) Hart, Borehole Kasatkin 7, depth interval of 2067–2074 m, Sample 115, prep. 7412, Tatarian Stage; (5) *Striomonosaccites* sp., Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 5, prep. 2838/1, Kazanian Stage; (6) *Corisaccites alutas* Venkatachala et Kar, Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 6, prep. 2522/1, Kazanian Stage; (7) *Vesicaspora schemeli* Klaus, Borehole Ivanovka 3, depth interval of 2364–2369 m, Sample 3671, prep. 2801/2, Tatarian Stage; (8) *Podocarpites* cf. *major* (Luber) Warjuchina, Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 5, prep. 2838/1, Kazanian Stage; (9) *Disaccites* sp., Borehole Ivanovka 1, depth interval of 3151–3154 m, Sample 2932, prep. 2477/2, Kazanian Stage; (10) *Striatopodocarpites* cf. *antiquus* Potonie, Borehole Ivanovka 3, depth interval of 2364–2369 m, Sample 6, prep. 2801/2, Tatarian Stage; (11) *Lueckisporites virkkiae* Potonie et Klaus, Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 6, prep. 2522/1, Kazanian Stage; (12) *Lueckisporites virkkiae* Potonie et Klaus, Borehole Ivanovka 3, depth interval of 2364–2369 m, Sample 3671, prep. 2801/2, Tatarian Stage; (13) *Taeniaesporites ortisei* Klaus, Borehole Ivanovka 3, depth interval of 3257–3166 m, Sample 6, prep. 2522/1, Kazanian Stage; (14) *Protohaploxypinus perfectus* (Naumova) Sauer, Borehole Kasatkin 7, depth interval of 2067–2074 m, Sample 115, prep. 7412, Tatarian Stage; (15) *Hamiapollenites tractiferinus* (Samoilovich) Jansonius, Borehole Ivanovka 3, depth interval of 3157–3166 m, Sample 6, prep. 2522/1, Kazanian Sage; (16) *Striatopodocarpites cancellatus* Clarke, Borehole Kasatkin 7, depth interval of 2067–2074 m, Sample 115, prep. 7412, Tatarian Stage; (17) *Hamiapollenites bullaeformis* (Samoilovich) Jansonius, Borehole Ivanovka 3, depth interval of 2364–2369 m, Sample 3671, prep. 2801/2, Tatarian Stage; (18, 19) *Vittatina flavata* (Maheschwari et Kar) Stanichnikova, ditto; (20) *Vittatina scutata* (Balme et Hennelly) Stanichnikova, ditto; (21) *Vittatina costabilis* Wilson, ditto.

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The Permian–Triassic Crisis in the Biosphere as Manifested in the Paleo-Pacific Deep-Water Sequences

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Abstract—The deep-water successions in Japan, Sikhote-Alin, and Nevada demonstrate a sharply decreased silica productivity across the Permian–Triassic boundary, accumulation of carbonaceous shales, and virtually absolute absence of fossil radiolarians and sponges. In Japanese sections, this stratigraphic level is enriched in titanium, phosphorus, and yttrium, and also demonstrate a sharp decline in cerium content relative to other rare-earth elements. These phenomena are directly or indirectly related to the biotic crisis at the Permian–Triassic boundary. The geochemical anomalies are interpreted as related to significant environmental changes in response to intense tectonic activity. The changes caused extinction of many marine biotic groups, including siliceous plankton, and increased the primary productivity of lower phytoplankton taxa. The event in question differs from the Mesozoic anoxic episodes in its origin, significance, and duration, being followed by the well-manifested biotic crisis.

Key words: Permian, Triassic, extinction, radiolarians, sponges, phytoplankton, biogenic silica production, geochemistry, carbon, rare-earth elements, anoxia, Paleo-Pacific.

The Phanerozoic biotic crises, including the most significant one at the Paleozoic–Mesozoic boundary, are under intense study at present. According to available data, the last one affected all groups of marine biota and resulted in the unprecedented mass extinction of genera and families. In addition to biotic events, different geochemical anomalies that have been considered as related to the crisis are recognized near the Permian–Triassic boundary. Up to the present time, investigations were focused mainly on the Paleozoic–Mesozoic boundary shallow-water sections. Their results are now essentially supplemented with data on the concurrent deep-water deposits that is discussed below.

Upper Paleozoic and Mesozoic siliceous deposits are widespread in all structures framing the Pacific. Nevertheless, siliceous beds across the Permian–Triassic boundary have been recognized only recently, and study on them just have been started. Results of the study are greatly valuable first for understanding of deep-water sedimentation during the Paleozoic–Mesozoic transition, which is absolutely unknown as yet, and second, for a new interpretation of causes and character of the Late Paleozoic crisis in the biosphere. The Permian–Triassic boundary deposits of deep-water origin are known in few land areas: Sikhote-Alin, Japan, South China, British Columbia, and Nevada. However, the sections studied show closely similar biostratigraphic and lithological features suggesting a uniform and regular structure of this stratigraphic interval in the Paleo-Pacific deep-water sections.

In the Sikhote-Alin, the most complete Lower Triassic deep-water section is known near the Dal'negorsk Settlement (Figs. 1 and 2). It characterizes the lower part of the Triassic–Lower Jurassic reference section of siliceous deposits in eastern Russia (Bragin *et al.*, 1988; Bragin, 1991, 1993). The Triassic deposits of the section are thrust over the Lower Cretaceous sandstones and are represented from the base upward by the following beds (the modified description after Bragin *et al.*, 1988):

(1) Black and dark gray carbonaceous shales with interbeds of gray siliceous mudstones (apparent thickness 3 m).

(2) Siliceous mudstones light bluish gray when fresh and rusty-brown to black when weathered. The rocks are massive, schistose, and rich in sulfides. In the upper part of the bed, there are few interbeds of black schistose carbonaceous mudstone and light gray chert (5 m).

(3) Light gray flaggy and dense, weakly siliceous mudstone with interbeds (1 or 2 cm) of light gray glassy cherts translucent and bearing conodonts *Neospathodus waageni* Sweet and *N. pakistanensis* Sweet of the early Olenekian (Smithian) age. The chert interbeds also contain abundant siliceous sponge spicules and single spherical specimens of undeterminable radiolarians (5 m).

(4) Brick- to crimson-red and black carbonaceous–siliceous mudstone intercalated with interbeds of light gray translucent chert. The bed contains conodonts *Neospathodus triangularis* (Bender) and *N. homeri*

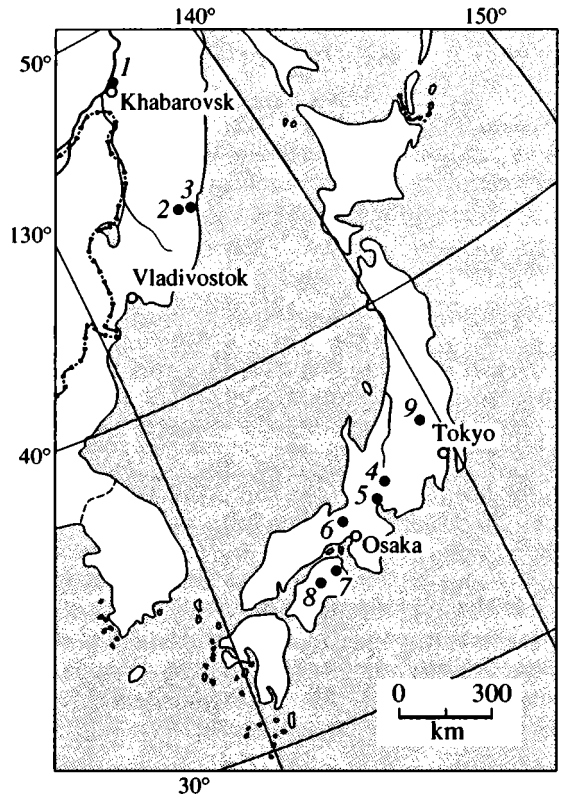


Fig. 1. Main Lower Triassic deep-water sections in Sikhote-Alin and Japan: (*1*) Khabarovsk; (*2*) Dal'negorsk; (*3*) Sadovyi; (*4*) Inuyama; (*5*) Kinkazan; (*6*) Sasayama; (*7*) Tenjinmaru; (*8*) Nakaoi; (*9*) Motegi.

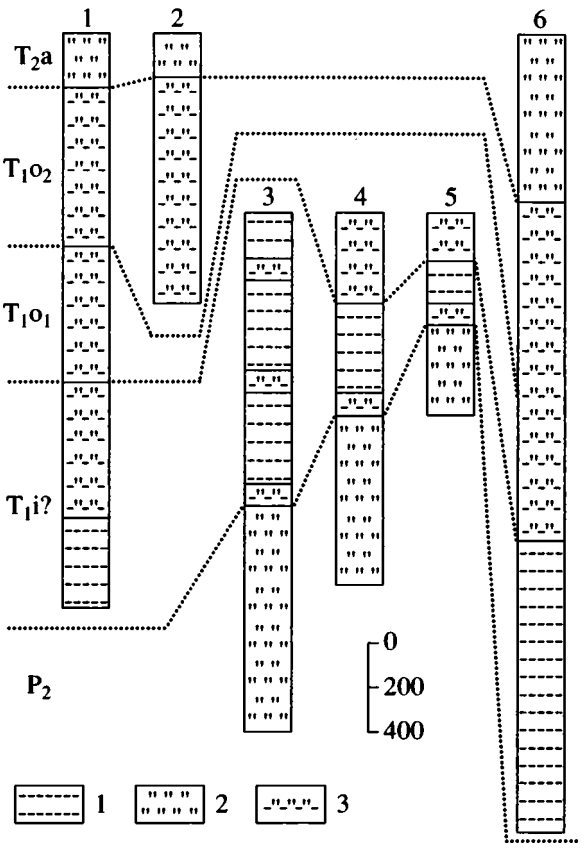


Fig. 2. Succession and correlation of the main Upper Permian–Lower Triassic deep-water deposits of Sikhote-Alin and Japan: (1) carbonaceous shale; (2) chert; (3) siliceous–clayey deposits.
(1) Dal'negorsk; (2) Sadovyi; (3) Nakaoi; (4) Tenjinmaru; (5) Sasayama; (6) Kinkazan. Symbols for stratigraphic units: P_2, Upper Permian; T_1i? presumable Lower Triassic, Induan Stage; T_1o_1, Lower Triassic, lower Olenekian Stage; T_1o_2, Lower Triassic, upper Olenekian Stage; T_2a, Middle Triassic, Anisian Stage.

(Bender) characteristic of the Upper Olenekian (Spathian) Stage and radiolarians *Cryptostephanidium longispinosum* (Sashida), *Pseudostylosphaera fragilis* (Bragin), *P.* sp., and *Parasepsagon* sp. cf. *leptaleus* Sugyama. Conodonts coexist with reworked Late Permian radiolarians *Follicuculus* sp. and siliceous sponge spicules (5 m).

(5) Gray siliceous mudstone with rare intercalations of light gray translucent cherts. Conodonts *Neospathodus homeri* (Bender) and *N. spathi* Sweet of the upper Olenekian (upper Spathian) Stage and radiolarians *Cryptostephanidium longispinosum* (Sashida), *Pseudostylosphaera fragilis* (Bragin), *P.* sp., *Parasepsagon* sp. cf. *Leptaleus* Sugiyama, and *Oertlispongus* sp. are detected in the bed (2 m).

Above the beds, there are lower Anisian deposits represented by flaggy cherts with subordinate interbeds of siliceous mudstones. The deposits contain abundant radiolarians characteristic of the entire *Hozmadia gifuensia* Zone (Bragin, 1991; Sugiyama, 1992).

The Lower Triassic deep-water deposits in other Sikhote-Alin areas demonstrate similar composition. For example, to the southeast of Dal'negorsk, near the Sadovyi Settlement (Bragin, 1991), the basal part of the section is composed of gray and dark gray siliceous mudstones up to 10 m thick. The mudstones yield the late Olenekian (Spathian) conodonts *Neospathodus triangularis* (Bender) (Figs. 1 and 2). In the Khabarovsk region, on the right bank of the Amur River near the railway bridge (Bragin, 1991, 1992), the Lower Triassic deposits in a tectonic block are presumably 20 m thick (the thickness may be overestimated because of folding) and represented by alternating gray beds of siliceous mudstone and clayey chert bearing the late Olenekian (Spathian) conodonts *Neospathodus triangularis* (Bender), *N. homeri* (Bender), and *N. spathi* Sweet (Fig. 1). According to lithological and biostratigraphic features, these deposits can be correlated to beds 2 and 3 of the Dal'negorsk section. Similarly to the Dal'negorsk section, the Khabarovsk and Sadovyi sections predominantly consist of clayey Lower Triassic deposits overlain by the Anisian radiolarian cherts.

Several Upper Permian–Lower Triassic deep-water sections are known in Japan. In the Shikoku Island, the

Nakaoi section (Kakuwa, 1996a, 1996b) exhibits the following beds (Figs. 1 and 2):

(1) Gray flaggy chert with radiolarians (*Neoalbaillella optima*) characteristic of the Dzhulfian Stage and lower Dorashamian Substage (9 m).

(2) Gray clayey chert of the upper Dorashamian Substage with conodonts *Neogondolella chanxingensis* and *N. subcarinata* (0.5 m).

(3) Black carbonaceous shale with horizons of gray siliceous mudstone (up to 15 m thick).

Bed 3 of the section is supposed to be of Lower Triassic (Induan) age, but paleontological remains are missing here. The bed is similar in lithology to the basal carbonaceous deposits of the Dal'negorsk section.

In the Tenjinmaru section of the eastern Shikoku Island (Kakuwa, 1996a, 1996b), the following deposits are described (Figs. 1 and 2):

(1) Gray flaggy cherts with radiolarians *Neoalbaillella optima* of the Dzhulfian Stage and lower Dorashamian Substage (8 m).

(2) Gray chert with intercalations of gray mudstone bearing conodonts *Neogondolella changxingensis* and *N. subcarinata* of the upper Dorashamian Substage (1.5 m).

(3) Black carbonaceous shale (1.5 m).

(4) Black carbonaceous shale with interbeds of gray dense siliceous mudstone (1.5 m).

(5) Gray massive siliceous mudstone with rare interbeds of gray chert; the rocks yield conodonts *Neospathodus waageni* Sweet, *N. dieneri* Sweet, and *N. conservativus* Sweet peculiar to the lower Olenekian (lower Smithian) Substage (4 m).

The data suggests that the member of carbonaceous shales (Bed 3 of the Nakaoi section and beds 3 and 4 of the Tenjinmaru section) corresponds to the stratigraphic interval from the uppermost Dorashamian to the lowermost Olenekian. Most likely it is of the Induan (Griesbachian–Dienerian) age. The lower Olenekian deposits (Bed 5) of the Tenjinmaru section are similar in lithology to coeval deposits of the Dal'negorsk section.

In the Honshu Island, analogous sections are known to the north of Osaka (the Sasayama section), in the Gifu area (the Kinkazan and Inuyama sections), and to the north of Tokyo in the Ashio Mountains (the Motegi section). In the Sasayama section (Ishiga *et al.*, 1996), the 5-m-thick bed of gray Dzhulfian–Dorashamian chert is overlain by the following succession attributed to the Early Triassic (Figs. 1 and 2):

(1) Dark gray siliceous mudstone (0.5 m).

(2) Black carbonaceous mudstones (0.8 m).

(3) Mudstone with intercalations of siliceous mudstone (1 m).

Permian deposits are missing in the Kinkazan section (Sugiyama, 1992), where the bed succession (from the base upward) is as follows (Figs. 1 and 2):

(1) Black carbonaceous shale (up to 20 m thick).

(2) Dark gray and gray siliceous-clayey shales (up to 15 m thick).

(3) Dark gray and black clayey cherts with abundant redeposited clasts of Permian cherts ranging in size from meter-scale blocks to microinclusions (up to 10 m).

Bed 1 contains no fossils other than single Permian radiolarians, which are evidently reworked. The upper part of Bed 2 and the lower part of Bed 3 yielded the Early Triassic (late Olenekian) assemblage of conodonts and radiolarians. The upper part of Bed 3 contains the early Anisian microfauna; Bed 4 is of the middle Anisian age (Sugiyama, 1992). The section thickness is likely overestimated because of tectonic doubling in some members.

In the Inuyama section (Fig. 1) accepted as the reference section of the Triassic–Jurassic deep-water deposits of Japan (Matsuoka *et al.*, 1994), the Lower Triassic part is represented by greenish gray siliceous–clayey shales with frequent interbeds of black carbonaceous mudstone and light gray dolomite; the total apparent thickness is 10–15 m. These deposits contain Olenekian conodonts *Neospathodus waageni* Sweet and *N. homeri* (Bender). Above, there are Anisian radiolarian cherts. Lithological composition of all rock types in the section has been thoroughly studied (Matsuoka *et al.*, 1994). The siliceous–clayey shales are composed of quartz microgranules and clay minerals. The black mudstones are enriched in carbon and carbonates. The dolomites consist of rhombohedral dolomite crystals also dispersed in other rock types. Similar deposits are widespread in adjacent areas of Japan (the Mino, Chichibu, and Northern Kitakami belts), where they are known as the facies of the "Toishi-type Shale" (Imoto, 1984; Ishida *et al.*, 1992; Sashida *et al.*, 1992; Saito, 1993). On the basis of conodonts, these deposits are attributed to the Olenekian (Smithian and Spathian) Stage (Koike, 1981). They are well correlated with coeval deep-water deposits of the Sikhote-Alin, to be exact, with beds 1–3 of the Dal'negorsk section.

In the Motegi section (Fig. 1), the Upper Permian gray cherts are overlain by black carbonaceous shales (up to 2 m) of the lowermost Triassic (?) and by gray siliceous mudstones (1 m) presumably of the Olenekian age, which are similar in lithology to "the Toishi type Shale" (Kamata and Kajiwara, 1996). The Motegi section shows an insignificant thickness of the Lower Triassic deposits.

In the East Pacific framing structures, deep-water deposits of the Lower Triassic were recognized in the Quinn River section, Nevada (Blome and Reed, 1995). The section begins with Upper Permian deposits. The Dzhulfian Stage (radiolarians of the *Neoalbaillella optima* Zone) is represented by flaggy cherts. The Dorashamian Stage (radiolarians of the *Neoalbaillella ornithoformis* Zone) corresponds to mudstone beds with rare horizons of black carbonaceous shales. The Induan Stage has not been recognized in the Quinn

River section. The presumable lower Olenekian Stage is represented by carbonaceous shales alternating with mudstones devoid of fossils. The upper Olenekian deposits with radiolarians *Parentactinia nakatsugawaensis* and the lower Anisian deposits with radiolarians *Hozmedia gifuensis* form the interval of alternating beds of siliceous mudstone and chert. Sections of similar type are known in the British Columbia (Isozaki, 1997) and were recently found in southern China (A. Yao, personal communication).

Other sections of Lower Triassic deep-water deposits have not been found as yet. Nevertheless, the available material is sufficient for outlining some regular compositional features of the Paleo-Pacific deep-water sections: (1) absence of the lowermost Triassic siliceous deposits; (2) gradual replacement of terrigenous deposits by siliceous upper Olenekian–lower Anisian sediments; and (3) abundance of carbonaceous shales in the lowermost Triassic interval, which become rapidly thinning upward and completely disappear in the lowermost Middle Triassic. These features are typical of all the sections described and, undoubtedly, characterize the history of the Paleo-Pacific deep-water sedimentation that was closely related to the global biotic and abiotic events at the Late Permian–Early Triassic transition.

It is very significant that the biogenic silica production ceased or noticeably diminished in the Early Triassic. In Japan, Sikhote-Alin, and many other regions, the Permian and Middle–Upper Triassic deep-water deposits are represented by facies of radiolarian cherts showing composition and structure persistently uniform in lateral directions and accumulated during a long period of time. The facies turned out to be untypical of the Early Triassic time, and absence of relevant beds in the sections is not associated with signs of regression and increased supply of clastic material, or with other indications suggesting changes in the bathymetry and morphology of sedimentary basins. The phenomenon can be reasoned only in terms of biotic crisis and degradation of siliceous plankton (radiolarians) that resulted in a sharp decline of biogenic silica production.

There are many facts confirming this inference. First, the biogenic origin of radiolarian cherts was established by many investigators (Vishnevskaya, 1984; Volokhin, 1985; Kaleda, 1987). In particular, the biogenic origin of SiO_2 of the Middle Triassic–Lower Jurassic siliceous deposits of the Inuyama section was revealed by the chemical and X-ray structural analyses (Matsuoka *et al.*, 1994). Second, in Japan and Sikhote-Alin, black shales of the presumable Induan age and siliceous mudstones of the lower Olenekian Substage contain no radiolarians other than the reworked Permian forms and similar spherical forms, which have not been identified as yet and probably belong to the order Entactinaria (Bragin, 1991; Sugiyama, 1992). Finally, sections of shallower deposits in different regions of the world are also characterized by the extremely poor radiolarian assemblages of the Early Triassic age. The oldest Triassic radiolarians are known from the upper Induan–lower Olenekian (Dienerian–lower Smithian) deposits of northwestern Turkey. They were found in dark gray and black limestones in association with conodonts, such as the late Induan *Neospathodus dieneri* Sweet and *Clarkina procerocarinata* Kozur, and the latest Induan–early Olenekian *N. dieneri* Sweet and *N. conservativus* (Muller) (Kozur *et al.*, 1996). Only three radiolarian species represented by *Entactinia (Stigmosphaerostylus) turkensis* (Kozur, Kaya et Mostler) and simple spherical spumellarians were found in these rocks. Younger (Late Olenekian or Spathian) radiolarian assemblages known from Japan, Thailand, and Sikhote-Alin also show the low diversity and predominance of primitive Entactinaria forms. Rare representatives of Mesozoic Nassellaria and Spumellaria are represented by primitive species in this case (Sashida, 1983, 1991; Bragin, 1991; Sugiyama, 1992; Sashida and Igo, 1992; Nagai and Mizutani, 1993; Kamata, 1995). All characteristic Permian taxa with advanced morphology are already lacking in the Lower Triassic deposits (Bragin, 1991; Kozur *et al.*, 1996).

Thus, many radiolarian taxa became extinct, and radiolarian plankton degraded in the Early Triassic time that is manifested as cessation of deep-water biogenic silica accumulation in the Paleo-Pacific region. To better elucidate this phenomenon, it is necessary to investigate in detail the eastern Mediterranean foldbelt (Tibet, southeastern Asia), where the Permian and Middle–Upper Triassic siliceous deposits are well known, but Lower Triassic beds have not been recognized. According to the taxonomic composition, the known radiolarian assemblage from Tibet that was originally dated back to Early Triassic (Feng, 1992; Feng and Liu, 1993) is, in fact, of the Middle Triassic (late Anisian–early Ladinian) age (Kozur *et al.*, 1996). As far as the Mediterranean region proper is concerned, the biogenic silica accumulation of the Permian–Early Triassic time is not recorded there, and sediments of the deepest settings are represented by clays and micritic limestones (Kozur, 1993, 1995a, 1995b).

The facts presented above motivate reconsideration of one of sedimentological problems, namely, the relation between biogenic silica accumulation and volcanism. Previous attempts were aimed to synchronize episodes of biogenic silica accumulation and volcanic activity (Zhamoida, 1992). The Permian and Triassic are known as periods of worldwide active volcanism of the trap-type in platforms and rift-related belts and adjacent areas. However, the biogenic silica accumulation did not proceed throughout the Early Triassic. The discontinuation in silica accumulation cannot be evidently correlated with the secular variations in volcanic activity and was certainly associated with other factors. This is an important argument in favor of genetic disconnection of volcanism and silica accumulation, as stated earlier (Bragin, 1991).

The dynamics of cessation and recommencement of biogenic silica accumulation is of great interest. The studied sections show that both events were not instantaneous or extraordinary. On the contrary, they were intermittent and gradual. For instance, the intensity of Late Permian silica accumulation in the Tenjinmaru and Nakaoi sections was analyzed using the so-called Ishiga diagram (Kakuwa, 1996a), where abscissa depicts the cumulative thickness of shale or mudstone interbeds in the siliceous sequence, and ordinate shows difference in the thickness of particular siliceous bed and the average thickness of cherty beds (Ishiga, 1994a). The analysis showed that the silica accumulation in the Tenjinmaru and Nakaoi sections decreased in the late Dzhulfian time and grew again in the Dorashamian age. The Upper Permian section of Central Japan (the Mino and Tamba belts) also demonstrates the reduction of silica accumulation and replacement of radiolarian siliceous deposits by mudstones within the upper part of the Dzhulfian Stage (Kuwahara and Ezaki, 1994). Thus, the late Dzhulfian time marks the first episode of diminishing and short-time interruption of silica accumulation prior to its complete cessation in the early Triassic.

The cease of biogenic silica accumulation in the Early Triassic was evidently not instantaneous. In the Sasayama section of southwestern Japan, par example, black carbonaceous mudstones with low SiO_2 content overlie a thin horizon of gray siliceous mudstones (Ishida *et al.*, 1992; Ishiga, 1994b) thus evidencing gradual changes in sedimentation. In many sections, black mudstone beds occur at the base of the Lower Triassic sequence and are devoid of siliceous interbeds or siliceous fossils, except the scarce and small spherical forms conventionally attributed to radiolarians (Kozur *et al.*, 1996).

In the early Olenekian time, the deposition of black carbonaceous mudstones was succeeded by accumulation of siliceous mudstones known in Japan (the Toishi-type Shale) and Sikhote-Alin. The lower Olenekian deposits of the Dal'negorsk section include numerous, though thin interbeds of flaggy cherts (Bragin, 1991). However, these cherts do not correspond to true radiolarites, because they contain only rare spherical remains of these organisms, but bear abundant spicules of siliceous sponges, which were evidently the rock-forming components.

Being abundant but not diversity, the spicules represented only Hexactinellida megascleres of the order Lyssakinosa (Plate). Skeleton of these sponges consisted of disconnected megascleres (Rigby, 1995). Pentactines, including those with straight and curved paratangentl rays, are most common, while hexactines and stauractines are less frequent.

This sponge assemblage is similar in composition to assemblages from the upper Induan–lower Olenekian strata of Turkey (Kozur *et al.*, 1996) and from the upper Olenekian beds of western United States (Rigby and Ganey, 1983). The extremely low diversity of the oldest Mesozoic sponge assemblages can be explained by the crisis experienced by these organisms also at the Permian–Triassic boundary. The lowest, highly siliceous interbeds of the Triassic sections in the Sikhote-Alin and Japan were formed owing to the biogenic silica production by sponges, whose life activity begun before the restoration of radiolarian plankton productivity.

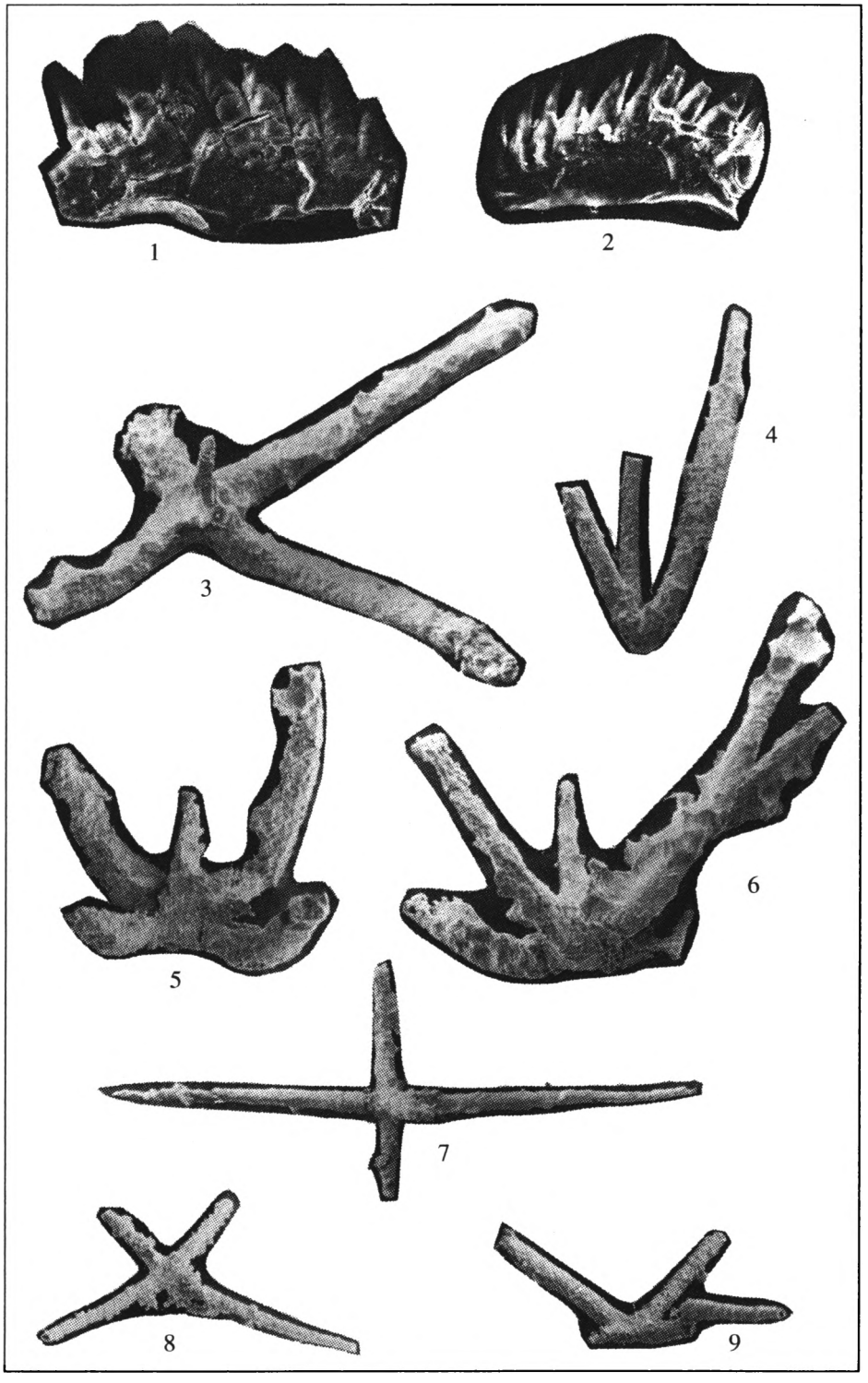
The Upper Olenekian deposits in Sikhote-Alin and Japan are also dominated by siliceous mudstones having low silica content and intercalated with frequent highly siliceous interbeds containing abundant spicules of siliceous sponges in association with radiolarian remains. True radiolarites occur locally. They become predominant higher in the section. The recommencement of biogenic silica accumulation that was interrupted in the earliest Triassic was a long gradual process. First signs of the true biogenic silica accumulation are confined to the basal Olenekian deposits in Sikhote-Alin and Japan, where siliceous components were produced by sponges (Kakuwa, 1996b). Radiolarians became again the rock-forming components in the late Olenekian time. The complete restoration of biogenic silica accumulation was in the early Middle Triassic.

The mentioned facts indicate that the discontinuation in the biogenic silica accumulation in the Paleo-Pacific deep-water areas was a manifestation of the crisis in marine biological systems during the Late Permian–Early Triassic transitional time. In addition to the absence of siliceous sediments, the sections studied demonstrate other records of this biotic crisis, which are equally remarkable.

First of all, this is the occurrence of the Lower Triassic sediments rich in carbon. They represent persistent markers at the bases of all known Lower Triassic deep-water sections. Some carbonaceous interbeds can be also found higher in the sequence, for example, inside the upper Olenekian interval of the Dal'negorsk section. However, the interbeds are untypical of the Permian, Middle and Upper Triassic deposits of Japan and Sikhote-Alin.

There is another, lower Toarcian level of carbonaceous deposits in the Mesozoic siliceous successions of these regions. Study on the Lower Jurassic siliceous deposits in Japan revealed that this level corresponded to the global anoxic event (Jenkyns, 1988; Hori, 1990, 1992, 1993; Isozaki and Maruyama, 1992; Ishiga, 1993; Matsuoka *et al.*, 1994). The lower Toarcian carbonaceous deposits widespread in Japan and Sikhote-Alin (Bragin, 1993) are of siliceous type and yield abundant and diverse radiolarian assemblages: they also bear scattered pyrite crystals and concretions. Reworked Triassic conodonts were found at this level and immediately above it in different sections (Bragin, 1993; Matsuoka *et al.*, 1994).

Similarity and difference between the Lower Triassic and lower Toarcian carbonaceous deposits were



Characteristic fossils of the Olenekian Stage, the Dal'negorsk section
(1, 2) conodonts: (1) *Neospathodus waageni* Sweet, (2) *Neospathodus homeri* (Bender); (3–9) megascleres of siliceous sponges of the Lyssakinosa Order: (3, 5–7) pentactines, (4) unidentified fragment of a large megasclere, (8, 9) tetractines.
Magnification is ×90 for all specimens.
Geological Institute of Russian Academy of Sciences, Coll. 4738.

revealed by their comparative chemical analysis. The lower Toarcian cherts contain 0.5–2% of free carbon, and the C/N ratio varies here from 7 to 20 that is supposed to be a result of carbon generation by the metabolic activity of phytoplankton, whose remains were found in these deposits (Ishiga, 1993). The Lower Triassic carbonaceous mudstones are characterized by similar C/N isotopic ratio but the free carbon content in these rocks is much higher (2% and more) (Ishiga, 1993). Because the rocks do not yield paleontological remains, their carbon was suggested to be of the bacterial genesis. The suggestion is supported by the composition of hydrocarbons and free kerogen (Ishiga, 1993; Suzuki *et al.*, 1993).

Genesis of the Early Triassic carbonaceous deposition is a subject of discussions. Some researchers consider anoxic environments (Matsuoka *et al.*, 1994; Kakuwa, 1996a, 1996b; Kamata and Kajiwara, 1996) or even the "superanoxia" of the Early Triassic time (Issozaki, 1994, 1997) as controlling events. The anoxic environments are thought to be the cause of the mass extinction at the Permian–Triassic boundary (Wignall and Hallam, 1992, 1993; Wignall and Twitchett, 1996). Other researchers suppose that the Lower Triassic carbonaceous deposits accumulated in response to the disturbance of the oceanic water stratification and active upwelling of oxygen-deficient bottom water masses enriched in biologically active compounds (Kajiwara *et al.*, 1994; Ishiga *et al.*, 1996; Kakuwa, 1996a, 1996b).

There is an evident and significant difference between the supposed "superanoxia" of the Permian–Triassic transition time and other Mesozoic anoxic events, for example, of the early Toarcian or Cenomanian–Turonian interval. Mass extinctions like that at the Late Permian–Early Triassic boundary are untypical of the early Toarcian and Cenomanian–Turonian events. As far as radiolarians are concerned, their evolution during and immediately after the anoxic events was characterized by some interesting and surprising phenomena, for instance, by mass appearance of new taxa without any noticeable extinction of others (Kuhnt *et al.*, 1986; Thurow and Kuhnt, 1986; Hori, 1992; Matsuoka *et al.*, 1994; O'Dogherty, 1994) and by a sharp increase in siliceous plankton productivity. For example, the upper Cenomanian and lower Turonian deposits of the Mediterranean region include widespread beds of radiolarian cherts, which are frequently carbonaceous (lydites) or alternating with carbonaceous strata. These are known as the "Bonarelli Horizon" of Italy (Marcucci-Passerini *et al.*, 1991) and its analogues in the Betic Cordillera of Spain (O'Dogherty, 1994), and as the "Ananuri Horizon" of the Caucasus. Thus, data on radiolarians and biogenic silica accumulation at the Permian–Triassic boundary are incompatible with data characterizing the early Toarcian and Cenomanian-Turonian anoxic events

It should be added that the period of carbonaceous sedimentation was much longer in the Early Triassic than in the early Toarcian or Cenomanian–Turonian. In my opinion, the Early Triassic event was actually a "superanoxia", as suggested by Isozaki (1993, 1994, 1997). In addition to the large scale, the Early Triassic anoxic event differs from other typical ones by some anoxia-related processes. Geochemical investigations of siliceous and other sections provide valuable information for understanding of the latter.

Geochemistry of the Upper Permian–Lower Triassic sediments was studied in different regions of the world. For example, the geochemical analysis of siliceous deposits has been started recently in Japan. In other countries, geochemical investigations were focused on carbonates of the relatively shallow genesis. As a result, significant geochemical changes were detected within this stratigraphic interval. These mutually correlatable changes correlative to other geological phenomena at the Paleozoic–Mesozoic boundary are of great significance for understanding the events of that time. The changes are manifested, first of all, in excursions of $\delta^{13}C_{carb}$, $\delta^{34}S$, and $^{87}Sr/^{86}Sr$ values.

The $\delta^{13}C_{carb}$ variations recorded in shallow-water carbonate sequences of the Alpine zone (Holser *et al.*, 1989, 1991) and Greenland (Oberhansli *et al.*, 1989) are very demonstrative. The $\delta^{13}C_{carb}$ value of 3.5‰ characterizing the Midian Stage (250 Ma) decreases to 2‰ in the upper Dzhulfian deposits (248 Ma). There was also a short positive excursion up to 3–3.2‰ (the lower Dorashamian Substage, 247 Ma) followed by the negative excursion down to –1‰ in the upper part of the stage (the Permian–Triassic boundary beds, 245 Ma old), which was particularly remarkable and rapid. Another positive $\delta^{13}C$ shift up to 0.5‰ (at the time of 244.5 Ma) was in the Griesbachian Stage, but the value declined again down to –0.5‰ in the upper part of the stage. Only after this last negative excursion, the value of $\delta^{13}C_{carb}$ increased again up to the high mid-Upper Permian level. Thus, the $\delta^{13}C_{carb}$ variation curve reveals three intervals of low values corresponding to the upper Dzhulfian Substage, Permian–Triassic boundary beds, and upper Griesbachian Stage of the Lower Triassic. These negative $\delta^{13}C_{carb}$ excursions mark periods of the declined productivity of marine biota. The first two intervals of low values coincide with two significant episodes of mass extinctions in the Late Permian (Stanley and Yang, 1994), and the third one corresponds to the earliest Triassic time of the lowest biota diversity and, correspondingly, of the low bioproductivity. The data are consistent with the rate of silica accumulation in Japan during the Late Permian–Early Triassic, where the silica accumulation first decreased at the end of the Dzhulfian Age, then grew in the Dorashamian time, and finally ceased at the Permian–Triassic boundary time (Kakuwa, 1996a).

Thus, the negative $\delta^{13}C_{carb}$ anomalies in the Mediterranean carbonate sections and events of decrease and

cessation of silica accumulation in the Pacific belt appear to be synchronous and can be interpreted as results of the low total productivity of marine biota. It seems that the occurrence of carbonaceous deposits in the Lower Triassic deep-water settings of the Pacific region contradicts this interpretation. However, carbon of these deposits is supposed to be the product of most primitive groups of phytoplankton, i.e., of bacteria or lower algae (Ishiga, 1993; Suzuki *et al.*, 1993; Ishiga *et al.*, 1994). The high primary productivity of this primitive microflora may oppress other groups of marine biota (Kakuwa, 1996a, 1996b). This may be the reason for the low restoration rates of radiolarian and other marine biotic communities during the Early Triassic (Alekseev, 1989; Hallam, 1991).

Another problem concerns the low $\delta^{13}C_{carb}$ value (–1‰) in the Permian–Triassic boundary carbonates, which cannot be simply attributed to the decreased productivity of marine biota. There are two interpretations of this phenomenon: (1) oxidation of previously concentrated organic carbon as a result of extensive sea regression and emergence of shelf areas in the latest Permian (Holser and Magaritz, 1987; Baud *et al.*, 1989; Holser *et al.*, 1989, 1991); and (2) significant hydrological changes and upwelling of bottom water masses enriched in the light carbon isotope (Malkowskii *et al.*, 1989; Grotzinger and Knoll, 1995; Knoll *et al.*, 1996). Geochemical investigations of the Paleo-Pacific deep-water deposits can provide valuable information for solution of the problem.

These investigations were carried out in a series of Japanese sections. Geochemical results, when analyzed and compared to data on carbonaceous horizons accumulated during other Mesozoic anoxic events, for instance, during the early Toarcian or Cenomanian–Turonian ones, contribute much to understanding of biotic events near the Permian–Triassic boundary.

It was found out that the Lower Triassic mudstones of the Tamba and Chichibu groups are enriched in REE, U, and Ba being simultaneously depleted in Mn, Ca, and Sr (Musashito, 1993). Similar data were obtained by geochemical analysis of the Lower Triassic deposits of the Sasayama and Tenjinmaru sections, where they show, in addition, the high content of TiO_2, P_2O_5, Y, and some other components (Ishiga *et al.*, 1996).

TiO_2 is one of basic components of igneous rocks, and its high content in mudstones lacking pyroclastic admixture is suggested to be a result of the activated continental alkali–basaltic volcanism (Ishiga *et al.*, 1996). In opinion of cited authors, the inference is supported by high Zr and Hf concentrations in the studied deposits. The high content of P_2O_5 and Y is given similar explanation, because their concentrations in the continental basalts are much higher than those in MORB and tholeiites of oceanic islands (Ishiga *et al.*, 1996). The above characteristics indicate a more intensive transportation of continental material into the ocean. However, it should be noted that the high P_2O_5 content in the Lower Triassic sediments can have another cause, i.e., a significant influx of organic material.

There are two possible explanations of this phenomenon. The increased continental supply in the Early Triassic World Ocean may resulted either from the intensified geological activity, or from the sea transgression and more intense river runoff. A valuable indicator of these phenomena is the Ce content relative to other REE. As compared to the Pacific, the modern Atlantic Ocean is known to be characterized by higher Ce concentration that is caused by more active inflow of Ce-rich river waters (Musashino, 1990). In addition, the deep water masses are depleted in Ce relative to the surface water layer (Liu *et al.*, 1988). In the Sasayama and Tenjinmaru sections, the high Ce content is recorded in the Upper Permian deposits; it abruptly decreases in the Lower Triassic beds and gradually grows upward. This observation contradicts the assumption that the continental runoff into the Paleo-Pacific was intense. The assumption is also inconsistent with data on strontium isotope composition in carbonate sediments. In the Upper Permian deposits, the initial $^{87}Sr/^{86}Sr$ ratios rapidly lower down to the minimum (0.7070) near the Permian–Triassic boundary (Smalley *et al.*, 1994). This trend suggest the increasing proportion of mantle influx into the World Ocean.

Different models are suggested to explain all geochemical phenomena observed and to settle the contradictions. According to one of them (Ishiga *et al.*, 1996; Kakuwa, 1996), the tectonic activity in the Permian–Triassic transition time improved connections between the Tethys and Paleo-Pacific. The water mass mixing caused a sudden change in the previously passive and stable Paleo-Pacific hydrodynamics that resulted in an active upwelling of bottom water masses depleted in Ce and O, but enriched in biologically active matter and toxic elements (Kakuwa, 1996a, 1996b). This induced the crisis of marine biological systems, which is manifested in the increased primary productivity of lower phytoplankton and in degradation of other groups. It should be noted that this model does not take into account the problem of water mass balance, because the water masses of the Paleo-Pacific were much more voluminous than those of the Tethys. Moreover, the destruction of oceanic barriers occurred more than once during the Phanerozoic history, but never resulted in rearrangement of the oceanic water structure.

Similar explanation is suggested for the behavior of sulfur isotopic ratios in the Motegi section of central Japan (Kamata and Kajiwara, 1996). Based on high $\delta^{34}S$ values (from –6 to –8.5‰), the appearance of anoxic conditions during the Late Permian was suggested. Across the Permian–Triassic boundary, the $\delta^{34}S$ values abruptly decrease (down to –30‰), and, in opinion of cited authors, this evidences a short-term cessation of anoxic conditions in response to the intensive

mixing of bottom and surface water masses. Afterward, in the earliest Triassic, the new phase of anoxic conditions gradually increased $\delta^{34}S$ values up to –10‰ and lasted for a long period, till the early Middle Triassic. The long-term anoxic event in the Early Triassic may be a consequence of events that activated oceanic hydrodynamics in the Permian–Triassic boundary time (Kamata and Kajiwara, 1996). The primary reason of hydrodynamic changes is unclear in this model.

According to another model, the eustatic fluctuations are factors responsible for typical anoxic events of the Mesozoic time. The model was discussed in detail by Gavrilov and Kopaevich (1996), who considered the Cenomanian–Turonian event as the principal example. In the Crimean sections, the anoxia-related deposits are enriched in organic carbon and toxic metals, such as Ni, Mo, and Zn. In general, a high content of toxic elements is characteristics of different deposits formed under anoxic conditions (Yudovich and Ketris, 1991). The organic carbon rapidly accumulated owing to the high primary productivity of phytoplankton and significant inflow of organic matter, nitrogen, and phosphorus compounds from continents. This was a response to the rapid flooding of vast maritime plains originated during the preceding regression. The concentration of toxic elements in sediments was favored by the lowered oxidation potential and by absorption of elements in sheaths of dead algae.

The comparison of these two models clearly shows the principal difference between anoxic events confined to the Permian–Triassic boundary and other Mesozoic epochs. In the first case, anoxia was caused by the upwelling of oxygen-deficient water masses rather than by transgression and subsequent phytoplankton bloom. This can explain the fact that the carbonaceous shales, which were formed under anoxic conditions, occur above the Permian–Triassic boundary in the Paleo-Pacific sections (Ishiga *et al.*, 1996; Kamata and Kajiwara, 1996). In this case, the active development of phytoplankton in pelagic settings of the Paleo-Pacific began after the mass extinction event and was for a long time a factor that decelerated restoration of marine biological systems. Therefore the mass extinction at the Permian–Triassic boundary was of a much greater significance than the biotic changes of the early Toarcian or Cenomanian–Turonian times.

Nevertheless, all the models suggested are disputable, and there are no sufficient data to prefer one of them. The most obscure are primary causes of the described events, and this aspect of the problem should be investigated further.

CONCLUSIONS

In the Paleo-Pacific deep-water sections, the Permian–Triassic boundary interval is characterized by significant changes in the taxonomic composition of fossil assemblages, and in lithology and geochemistry of sediments. All these changes are directly connected with the most significant crisis in the biosphere that occurred near this boundary. Some changes are related to the causes of the crises, while others are its manifestations or consequences.

The negative Ce anomaly and positive anomalies of TiO_2, P_2O_5, and Y seem to be a consequence of the activate tectonic regime. This factor could provoke the biotic crisis.

The biotic crisis is manifested directly or indirectly in the sections studied. The direct manifestations are the extremely poor taxonomic composition of radiolarian assemblages and siliceous sponges. The indirect manifestations or consequences are the decreased rate of biogenic silica accumulation and widespread deposition of sediments enriched in organic matter.

The biotic and geochemical changes recorded in the sections are well correlative to each other and traceable in lateral directions (Figs. 1 and 2). On one hand, this suggests a connection between the biotic and abiotic phenomena, and, on the other, it is clear the integrated application of paleontological and geochemical data is useful for the stratigraphic scale refinement.

The large scale and duration of the Early Triassic crisis, along with its biotic, lithological and geochemical manifestations reveal its unique character in the history of the Earth and principal difference from other well known anoxic events.

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The Age of Chokusu Beds in the Kyzyl-Dzhar-Chokusu Section (the Silurian of Northwestern Mongolia)

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Abstract—The rock succession and age of the Chokusu Beds in the Mt. Kyzyl-Dzhar-Chokusu section is considered to be either the Ludlovian–Pridolian or Wenlockian–early Ludlovian. The discrepancy resulted from a complicated geological structure and contradictory interpretations of rock ages on the basis of different groups of invertebrates (tabulates, heliolitids, rugoses, bryozoans, brachiopods). We reckon the beds to be of the Wenlockian age. It is not inconceivable that the upper Llandoverian and lower Ludlovian deposits are also present in the section.

Key words: biostratigraphy, Silurian, Chokusu Beds, invertebrates, northwestern Mongolia.

Silurian terrigenous–carbonate deposits with abundant fauna of stromatoporoids, corals, bryozoans, brachiopods, and crinoids are exposed on the northwestern slope of Mt. Kyzyl-Dzhar-Chokusu in northwestern Mongolia, where they were attributed to the Chokusu Formation (Kopaevich and Ulitina, 1977) and later to the Chokusu Beds (Sytova and Ulitina, 1983).

As a result of the thirty-year study, two basic versions concerning subdivision, correlation, and age of the deposits, and also the block structure of the mount are suggested. According to the first version, the Chokusu Beds of Mt. Kyzyl-Dzhar-Chokusu are dated back to the Wenlockian–early Ludlovian, and, probably, to the late Llandoverian, whereas the second version implies the Ludlovian–Pridolian and probably Early Devonian age of the beds (Fig. 1).

Abundant fauna found in the deposits is of great importance for correlation, and the uncertainty in dating is by no means unimportant for paleontologic and geologic research.

In addition to our own field observations and results of comprehensive study of tabulates (Minzhin) and heliolitids (Bondarenko) presented in the work, data available in publications were analyzed as well. Minzhin took part in the field work together with Kh.S. Rozman (seasons of 1975, 1976, 1981, and 1983), and Bondarenko with L.M. Ulitina in 1977.

INVESTIGATION HISTORY OF THE CHOKUSU BEDS

The name Kyzyl-Dzhar-Chokusu[1] is given to both the steep slope of the mount (altitude 2701.8 m) and the creek at its foot. The name is of Uigur origin and means "Pure Water of Red Scarp;" in Mongolian, the mount is named Ulaan-Ereg, and this is the reason why different names can be encountered in papers and maps. The mount is situated in northwestern spurs of the Mongolian Altai, 150 km to the northwest of Ulan-Khus Somon (Bayan-Ulgii district) on the right bank of the Ikh-Oiguringol River in its middle courses. The site is close to the boundary between Mongolia, China, Kazakhstan, and Russia (Fig. 2a).

An extensive study on geology and fauna of Mongolia was initiated in 1963 in connection with the geologic mapping at the scales 1 : 1 500000 and 1 : 500000 (*Geologicheskaya karta...*, 1978). The Kyzyl-Dzhar-Chokusu rocks belong to the Silurian sequences characteristic of superimposed structures (Dergunov *et al.*, 1980). Summarizing data on Silurian stratigraphy of Mongolia, Filippova and Luvsandanzan (1973) stated that rocks exposed in the mount must be included in the undivided Wenlockian–basal Ludlovian sequence. The rocks were characterized in the summary as follows: "... A.B. Dergunov considers the variegated terrigenous sediments that fill up a small superimposed trough in the middle course of the Oigurin-Gol River as Silurian in age. These sediments rest with a sharp unconformity on the underlying Cambrian–Ordovician deposits and are overlain by the Middle Devonian volcano-sedimentary deposits. They are 600–800 m thick. Among brachiopods collected from the lower horizons of the sequence, R.E. Alekseeva identified *Stegerhynchus decemplicatus angaciensis* Tchern., *Camarotoechia ubsuensis* Tchern., *Tannuspirifer pedaschenkoi* Tchern., *Tuvaella rackovskii* Tcheru., *Spondylostrophia* ex gr. *lata* Kulkov, *Leptostrophia elegestica* Tch-

[1] Synonyms common in Russian publications are Kyzyl-Dzhar-Chokusu, Bayan-Ulegei, Ikhe-Oigurin-Gol, Ikh-Oiguryn-Gol, Oigurin-Gol, Oiguriin-Gol, and Oiguringol.

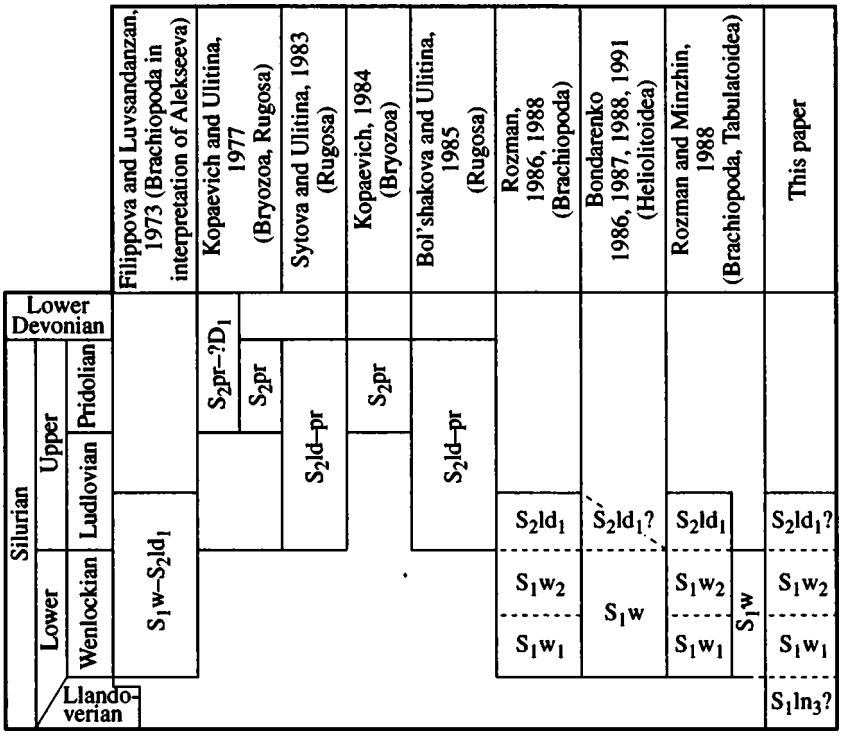

Fig. 1. Age of the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu estimated by different researchers.

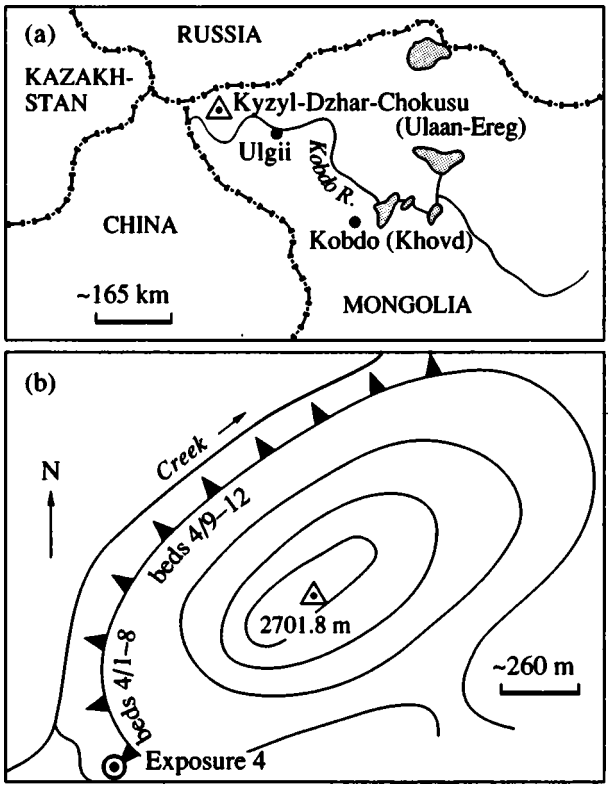

Fig. 2. Mount Kyzyl-Dzhar-Chokusu: (a) geographic position; (b) structure in topography and sequence of beds (4/1–8 and 4/9–12) with fauna (see also Fig. 3).

ern., *Howelella* sp., *Plectatrypa* sp. typical of the Wenlockian Stage of the Silurian System" (Filippova and Luvsandanzan, 1973, p. 151).

Another synthesis on Silurian rocks of Mongolia (Dergunov *et al.*, 1980) was published in the explanatory notes to the geologic map of Mongolian Altai, scale 1 : 500000 (1978). Fossils found in Silurian deposits of Mount Kyzyl-Dzhar-Chokusu are tabulates *Halysites bifiolus* Kovalevsky and *Favosites interstinctus* Regnell (identified by T.T. Sharkova); heliolitids *Helioplasma* sp. (identified by T.T. Sharkova); rugoses *Phaulactus* sp., *Carinophyllum confusum* (Pocta), and *Pilophyllum weissermeli* Wedekind (identified by V.A. Sytova); and brachiopods *Stegerhynchus decemplicatus angaciensis* Tchern., *Camarotoechia ubsuensis* Tchern., *Plectatrypa* sp., *Howelella* sp., *Tannuspirifer pedaschenkoi* Tchern., *Tuvaella* sp., and *Spendilostrophia* ex gr. *lata* Kulkov (identified by Kh.S. Rozman). This faunal assemblage "allows the assignment of enclosing deposits to the Wenlockian and lower Ludlovian" (Dergunov *et al.*, 1980, p. 48).

Participants of the Joint Soviet–Mongolian Paleontological (JSMPE) and Geological (JSMGE) expeditions from the USSR Academy of Sciences and Mongolian People's Republic, who collected fossils and studied geological structure of the Mt. Kyzyl-Dzhar-Chokusu since 1974 are L.M. Ulitina (1974, 1975, 1977), G.V. Kopaevich (1974), L.N. Bolshakova (1975), Kh.S. Rozman, Ch. Minzhin (1975, 1976,

1981, 1983), and O.B. Bondarenko (1977; indicated in parentheses are years of fossil collecting).

Kopaevich and Ulitina (1977) distinguished the 400-m-thick sedimentary sequence of the Kyzyl-Dzhar-Chokusu site as the Chokusu Formation and published its description. They determined the rock age on the basis of bryozoans and rugoses as "...the Late Silurian and probably Pridolian...," stating that "their assemblage can be typical of the Silurian–Devonian boundary layers," and that "... the composition of rugoses and bryozoans implies the Pridolian and probably earliest Gedinnian age of the Chokusu Formation" (p. 49, 60; see Fig. 1). The Chokusu Formation was subdivided into three lithological units: the lower, middle and upper. The paper by Kopaevich and Ulitina (1977) also describes the live forms of corals and bryozoans from the Chokusu Beds enabling reconstruction of their habitat environments. In addition, they showed stratigraphic ranges of other invertebrates (stromatoporoids, tabulates, heliolitids, and brachiopods) in the section, but without identification of their generic and species levels.

In their monograph dedicated to rugoses, Sytova and Ulitina (1983) ranked the Chokusu Formation as the Chokusu Beds and suggested on the basis of rugoses that the deposits are older as compared to their age determination in 1977. They attributed the rocks to the Ludlovian–Pridolian, pointing out that "...the Pridolian rugoses dominate," and that "...the attempts to subdivide the Ludlovian and Pridolian of Mongolia have not been successful" (Sytova and Ulitina, 1983, p. 28, 37). Preliminary identifications of rugoses (Kopaevich and Ulitina, 1977) were revised, and some species were renamed as follows (new names are given in brackets): *Carinophyllum confusum* (Pocta) [*Stereoxylodes clarus* Sytova], *Pseudamplexus fascicularis* Soshkina [*Ps. quadripartitus* Soshkina], *Tryplasma weissermeli* Wedekind [*T. aequabile* (Lonsdale)], *Scyphophyllum clavum* Strelnikov [*S. vulgare* Sytova], *Ramulophyllum heterozonale* Nikolaeva [*Scyphophyllum vulgare* Sytova], *Cystiphyllum* aff. *siluriense* Lonsdale [*C. omphyforme* Grabau], *Kodonophyllum* aff. *truncatum* (L.) [*K. forte* Ulitina], *Lamprophyllum* sp. [*Coronoruga idonea*], *Stortophyllum* sp. [*Pholidophyllum loveni longiseptatum* Sytova].

Later, Ulitina (1993) attributed species *Stereoxylodes clarus* Sytova to the genus *Entelophyllum* and considered *Scyphophyllum vulgare* Sytova as the type species of the genus *Emedophyllum*.

In a new monograph, Kopaevich (1984) revised the majority of her preliminary determinations (Kopaevich and Ulitina, 1977) and described most bryozoans as new species. She interpreted the age of the Chokusu Beds as the Pridolian.

In 1985, Bol'shakova and Ulitina were first to show the Mt. Kyzyl-Dzhar-Chokusu structure as representing a series of tectonic blocks (Fig. 3a). Despite this, they followed the geological interpretation of 1977 implying that the Chokusu Beds section 360–400 m thick begins at the southwestern edge of the mount to continue further northeastward. The beds were attributed to the Ludlovian–Pridolian (Bol'shakova and Ulitina, 1985, p. 19, 33).

Extensive studies on brachiopods (Rozman, 1986, 1988) and Silurian heliolitids of Mongolia (Bondarenko, 1986–1988, 1991) showed that the Chokusu Beds are older than it was proposed earlier (Kopaevich and Ulitina, 1977; Sytova and Ulitina, 1983). Rozman (1986, 1988) established the Wenlockian–earliest Ludlovian age on the basis of brachiopods, pointing out that Bed 14 corresponds to the early Wenlockian, and beds 25, 29, and 34a to the late Wenlockian (see Fig. 3b for bed numbers). Bondarenko (1986–1988, 1991, 1999) argued for the Wenlockian age of the Chokusu Beds on the basis of heliolitids leaving as debatable the early Ludlovian strata.

When studying the entire geological profile of the Mt. Kyzyl-Dzhar-Chokusu, Rozman and Minzhin (1988) confirmed its complicated blocky structure established by Ulitina and her colleagues, but they differently interpreted the succession of the Chokusu Beds in many cases. Moreover, Rozman and Minzhin (1988) included into the Chokusu Beds the rocks of blocks IX and VIII in addition to blocks A–G, the section of which was considered by Ulitina and her team as the stratotype (Fig. 3a, b). The revised section of the Chokusu Beds was attributed to the Wenlockian by Minzhin on the basis of tabulates and to the Wenlockian–early Ludlovian by Rozman on the basis of brachiopods. In accordance with lithologic composition and brachiopod distribution, the Chokusu Beds were subdivided into three members bearing brachiopod assemblages. Moreover, Rozman and Minzhin were first to show that the Chokusu Beds constituting only a part of the mount contact along the thrust fault with the Oigurin-Gol Beds exposed in another part of the mount (Fig. 3b, c). The thickness of the Chokusu Beds was defined as equal to 230 m (190 m of the Wenlockian and 40 m of the Ludlovian intervals).

These are two basic interpretations of geological structure of the Mt. Kyzyl-Dzhar-Chokusu and of the Chokusu Beds age in its lower part.

GEOLOGICAL STRUCTURE AND FAUNA OF THE CHOKUSU BEDS

The geological structure. The southeastern slope and partially the top of the northwestern slope of Mt. Kyzyl-Dzhar-Chokusu are gentle and covered with grass, whereas rocks are well exposed in the major part of the steep northwestern slope. The Chokusu Beds make up nearly a half of the mount, being exposed in blocks as fragments of different monoclinal structures (Fig. 2b). Dip angles of the Chokusu Beds in blocks I–VIII vary from 40° to 85–90° (55–75° in average); the beds strike mainly northeastward (5–30°), sometimes

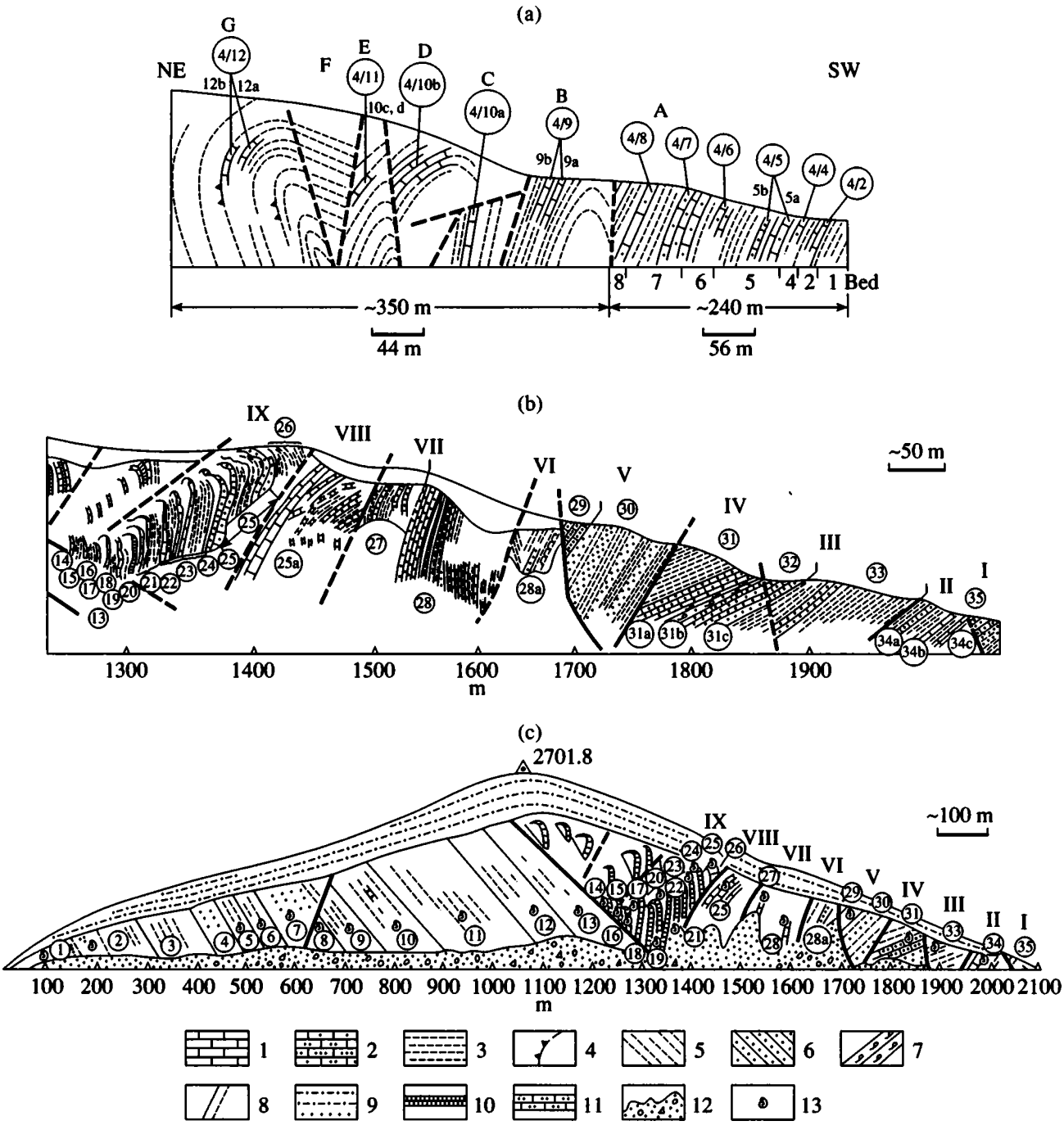

Fig. 3. Geological structure of the Mt. Kyzyl-Dzhar-Chokusu: (a) geological profile of the southern part of the mount after Bol'shakova and Ulitina (1985) with modifications by Bondarenko; (b) geological profile of the southern part of the mount after Rozman and Minzhin (1988); (c) geological profile of the mount made up by Minzhin after 1988.

Encircled numbers: (1–12) the Oigurin-Gol Beds; (14–35 and 4/2–4/12) the Chokusu Beds; (A–G, I–IX) indices of tectonic blocks. Symbols: (1) limestone; (2) sandy and clayey limestone; (3) siltstone with interlayers of mudstone and quartzite; (4) base of the bed; (5) siltstone and mudstone; (6) siltstone and sandstone; (7) sandy coquina; (8) faults; (9) siltstone and sandstone; (10) limestone; (11) sandy limestone; (12) talus; (13) fauna localities.

southwestward and northwestward (190–345°). Dip angles in Block IX, where the bed strike northwestward (350°) and northeastward (25°), are 50–90° in average. True relationships between the Chokusu Beds and underlying and overlying deposits are not established.

The Chokusu Beds represent a carbonate–terrigenous sequence of alternating siltstone, sandstone,

quartzite, mudstone, and limestone varying in color from light gray and greenish gray to brownish gray, yellow–brown and cherry. The rocks listed above dominate in various parts of the section. Limestones are sandy and clayey; locally, they are represented by detrital or biohermal, massive and layered varieties. Terrigenous rocks are sometimes enriched in carbonate material, and calcareous concretions can be found in them especially in clayey facies. All rocks mentioned above often replace each other along the strike. It is difficult to estimate thickness of the Chokusu Beds because of the complicated geological structure, but they appear to be as thick as 500 m. Below, we present a brief lithological characteristics of the Chokusu Beds in individual tectonic blocks. The correlation of blocks I–VII with blocks A–G and beds 27–34 with beds 4/2–4/12 is shown in Figs. 3 and 4.

Block IX includes two, carbonate and terrigenous members of the Chokusu Beds 130 m thick in total. The carbonate member 30 m thick (beds 14–21) is made up of light gray limestone with subordinate interlayers of greenish gray siltstone and mudstone; limestone is massive and layered, locally biohermal. The terrigenous member 100 m thick (beds 22–26) is composed of greenish gray siltstone and mudstone alternating in different ways with sandstone and sandy to clayey limestones varying in color from greenish gray to brownish gray.

In Block VIII (Bed 25a), the Chokusu Beds 50 m thick are represented by greenish to brownish gray limestone, sandstone, siltstone, and mudstone, which are displaying an intense cleavage and repeatedly replacing each other along the strike.

Block VII (beds 27, 28) consists of two, carbonate–terrigenous and terrigenous members, both up to 100 m thick. Color of rocks varies from rusty–brown to cherry, and, sometimes, to bluish and pinkish gray. The carbonate–terrigenous member 30 m thick (Bed 27) is represented by alternating sandstone, siltstone, mudstone, and calcareous sandstone intercalated with interlayers of clayey, sandy and detrital limestones; limy concretions occur in places in sandstone and siltstone. The terrigenous member 70 m thick (Bed 28) is made up of mudstone and siltstone with interlayers of sandstone and quartzite. Flyschoid rhythms encountered in places are useful for distinguishing the base and top of interlayers with hieroglyphs and graded bedding.

Block VI (Bed 28a) exhibits isolated outcrops of rocks similar to those of Block VII.

Block V (beds 29, 30) includes the 100-m-thick sequence of rocks similar to those of Block VII, but rusty brown and cherry-colored mudstone and siltstone are dominant rock types in this case.

Block IV (Bed 31) exhibits the 120-m-thick terrigenous sequence of sandstone, quartzite, sandy coquina, siltstone, and mudstone alternating with interlayers of calcareous sandstone and platy to grumous sandy and clayey limestones. The rock color is diverse, varying from greenish gray to rusty brown and cherry.

Bol'shakova and Ulitina, 1985			Rozman and Minzhin, 1988			
Age (Ulitina)	Block	Bed	Bed	Block	Age: Rozman	Age: Minzhin
pr	G	4/12	27, 28	VII	ld_1	w
pr	E	4/11, 4/10c, d	28a	VI	?	
pr	C, D	4/10a, b	29, 30	V	w_2	
pr	B	4/9	a b 31 c	IV	ld_1 w_2	
pr ld	A	4/8 4/7 4/6 4/5 4/4 4/3 4/2	32, 33 a b 34 c	III II	w_2 ld_1 w_2	

Fig. 4. Correlation of tectonic blocks and beds with fauna distinguished by Bol'shakova and Ulitina (1985), and by Rozman and Minzhin (1988).

Block III (beds 32, 33) consists of a carbonate–terrigenous sequence 100 m thick. Sandstone and siltstone layers are intercalated here with interlayers of quartzite, mudstone, calcareous sandstone, and sandy limestone. Limestone interlayers are mainly gray, with green, blue and yellow tint; cherry color is peculiar of Bed 32.

Block II (Bed 34) exhibits the 80-m-thick sequence of greenish gray siltstone and calcareous sandstone alternating with subordinate interlayers and bands of mudstone and concretions, and also with lentils of biohermal limestone.

Block I (Bed 35) consists of gray to bluish gray siltstone and quartzite devoid of fauna.

All the blocks, except Block I, yielded fossil invertebrates: corals, stromatoporoids, and bryozoans dominate in calcareous rocks, whereas brachiopods, which make up local coquina lentils, prevail in terrigenous rocks. The layer-by-layer, more comprehensive description of rocks of the Chokusu Beds was given by Kopaevich and Ulitina (1977), and by Bol'shakova and Ulitina (1983). Their brief characteristics were presented by Rozman and Minzhin (1988).

The alternating prevailing terrigenous, carbonate–terrigenous, and carbonate rocks, their mutual replacement along the strike, and presence of diverse corals indicate a shallow water regime of sedimentation in the Silurian basin. Despite the great variability of deposits, facies transitions along the strike, and controversial interpretations of the Chokusu Beds age in the Mt. Kyzyl-Dzhar-Chokusu section, the following inferences can be regarded as proved (Figs. 3, 4):

(1) The mount is of the blocky structure.

(2) Blocks A–G in the geological profile compiled by Ulitina and her team correspond to blocks I–VIII in the geological profile compiled by Rozman and Minzhin. Block IX was not studied by Ulitina's team.

(3) Tectonic blocks D and VI, C + D and V, A and I–III are similar in shape and appear to be the real structural elements. Researchers differently understand the geological age of the blocks and structural features of the mount.

The blocky structure of Mt. Kyzyl-Dzhar-Chokusu and steep dip angles of beds complicate the establishing of the true rock succession. The monoclinal bedding of rocks does not ensure the right succession of rocks, as the overturned bedding is possible. The suggested bed successions based on different groups of invertebrates are ambiguous and contradictory (Fig. 4).

TAXONOMIC COMPOSITION OF FOSSIL ASSEMBLAGES

The Chokusu Beds yield stromatoporoids, corals, cephalopods, bryozoans, brachiopods, and crinoids. Conodonts and graptolites have not been found so far, and crinoids have not been identified as yet. Two lists of tabulates are given in Figs. 5 (collections and identifications by Minzhin) and 6 (collections and identifications by other researchers). Heliolitids are listed again twice, in Figs. 5 (collected by Rozman and Minzhin, identified by Bondarenko) and 6 (collected and identified by Bondarenko). The taxonomic composition of fossils studied is analyzed below in terms of their assemblages and stratigraphic ("older–younger") relationships without subdivision into stages that will be discussed below.

The analysis of brachiopods and tabulates found in Block IX (Fig. 5) revealed two similar assemblages: the lower gipidulid and upper atrypid–leptaenid assemblages of *brachiopods* (Rozman and Minzhin, 1988) associated with corresponding assemblages of *tabulates*. Lower tabulates are represented by catenulate (*Halysites*) and tabular (*Syringopora*) forms. This tabulate assemblage almost lacks faveolate forms, except *Favosites jaaniensis* Sokolov. *Halysites* and *Syringopora* still exist in the second assemblage, but abundance and diversity of faveolate Tabulata (*Mesofavosites, Favosites, Barrandeolites*) increases here. *Heliolitids* are found in Block IX beginning from Bed 25, where *Helenolites clarus* Chekhovich and *Helioplasmolites dzharensis* Bondarenko are identified. Thus, the renewal of fauna begins from Bed 25 rather than from Bed 21, and, accordingly, we place the boundary between two assemblages higher than this was done by Rozman and Minzhin (1988).

Tabulates from the Chokusu Beds in blocks II, IV and Bed 27 of Block VII (Fig. 5) are represented by one assemblage of characteristic triad of species from orders Halysitida (*Halysites*), Syringoporida (*Syringopora*), and Favositida (*Palaeofavosites, Mesofavosites, Favosites, Tuvaelites, Barrandeolites*). Lists of tabulates from the Chokusu Beds, which were presented by Chudinova (1986), and by Bol'shakova and Ulitina (1985), correspond to the same triad, but the order Syringoporida is represented in their works by a common genus *Syringopora* (blocks A, B), though species are different (Fig. 6). In the order Halysitida, they identified the *Catenipora*, but not the *Halysites* genus (Block A), and genera *Thecostegites* and *Armalites* missing from our collections were determined in the order Favositida (none of particular block is pointed out). According to these data, three distinguished assemblages of tabulates are closely related to each other: the first is from beds 15–24 of Block IX, the second from Bed 25 of Block IX, and the third one from Bed 27 of Block VII, Bed 31 of Block IV, and Bed 34 of Block II. The first and second assemblages are characterized above. The third assemblage differs from them: it includes genera *Palaeofavosites* (two species) and *Tuvaelites* (one species), and has another species composition of genera *Halysites* and *Favosites*.

Heliolitids from the Chokusu Beds of blocks A–D are represented by two very similar assemblages comprising six common species of eight present and differing because of presence of the genus *Sytovaelites* in one of them. Taking into account Block IX, four heliolitid assemblages distinguished in the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu are one from Bed 25 of Block IX, second from beds 4/2–4/5 of Block A, third from Bed 27 of Block VII, Bed 31 of Block IV, and beds 4/6–4/9 of blocks A, B, and fourth from beds 4/10a and 10b of blocks C, D. The first assemblage consists of *Helioplasmolites dzharensis* Bondarenko only (Bondarenko, in press). The second assemblage includes many genera, but genera *Helioplasmolites* and *Sytovaelites* are missing in this case. The genus *Sytovaelites* appears in the third assemblage, and species *Diploepora vaga* Bondarenko is characteristic form of the fourth assemblage (Bondarenko, 1986, 1987). The first, second and third assemblages are successive in the section from the base upward. The stratigraphic level of the fourth assemblage that includes, in addition to species occurring in the second and third assemblages, the older (middle Llandoverian–Wenlockian) species *Diploepora vaga* Bondarenko is not clear. Rocks with the *Diploepora vaga* Bondarenko species (Bed 4/10) correspond to small calcareous bioherms, highly broken and displaying numerous niches and small cavities.

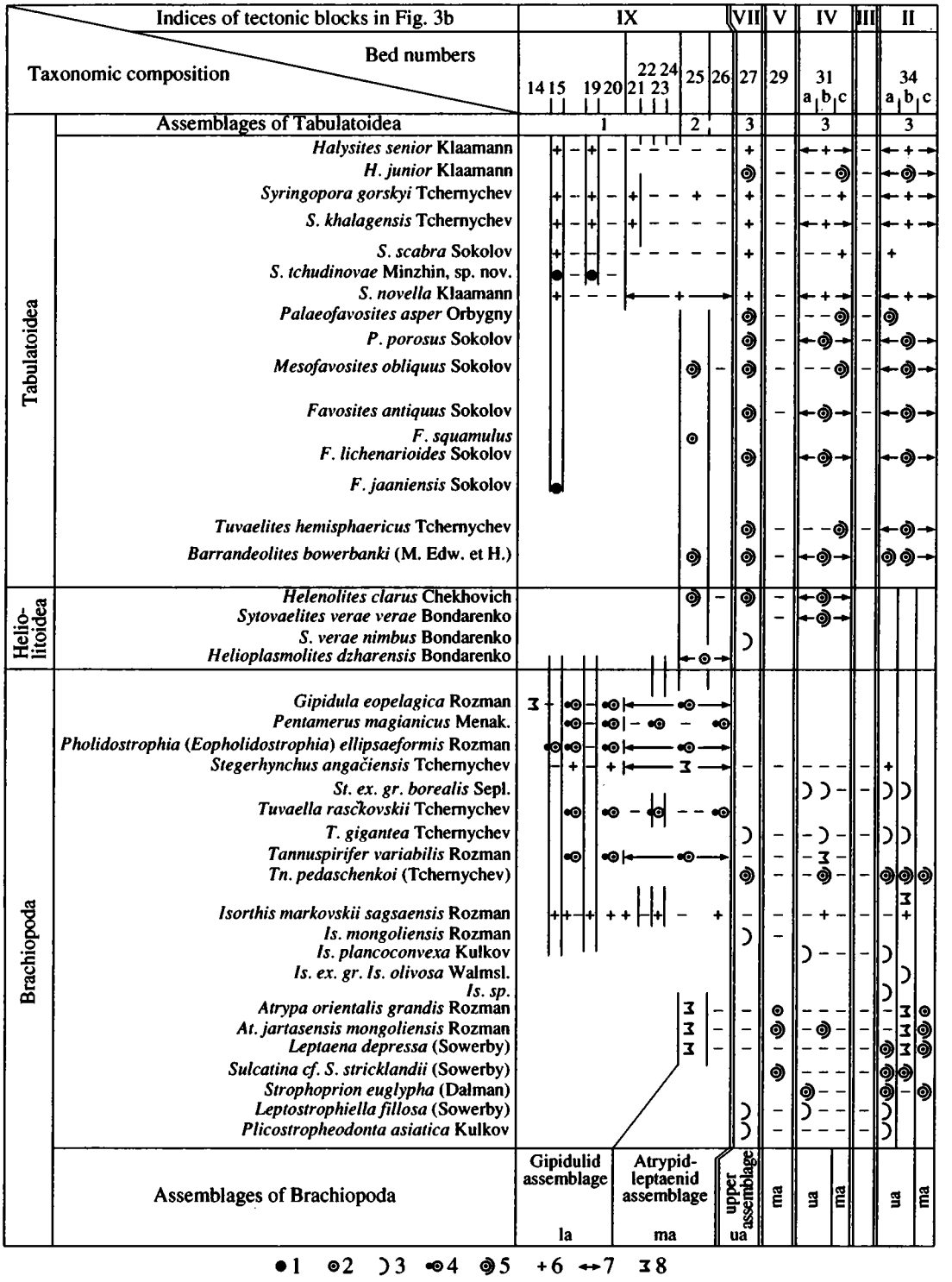

Fig. 5. Distribution of tabulates, heliolitids, and brachiopods in the Chokusu Beds, tectonic blocks II–IX of the Mt. Kyzyl-Dzhar-Chokusu (list of fauna is after Rozman, 1986, 1988; Rozman and Minzhin, 1988; Bondarenko, 1986, 1991, 1992, 1999; data of this work are included).

Faunal assemblages: (la) lower, (ma) middle, and (ua) upper (subdivision Rozman and Minzhin,1988). Species of the lower (1), middle (2), upper (3), lower and middle (4), middle and upper (5), lower, middle and upper (6) assemblages; (7) unspecified bed, (8) abundant specimens.

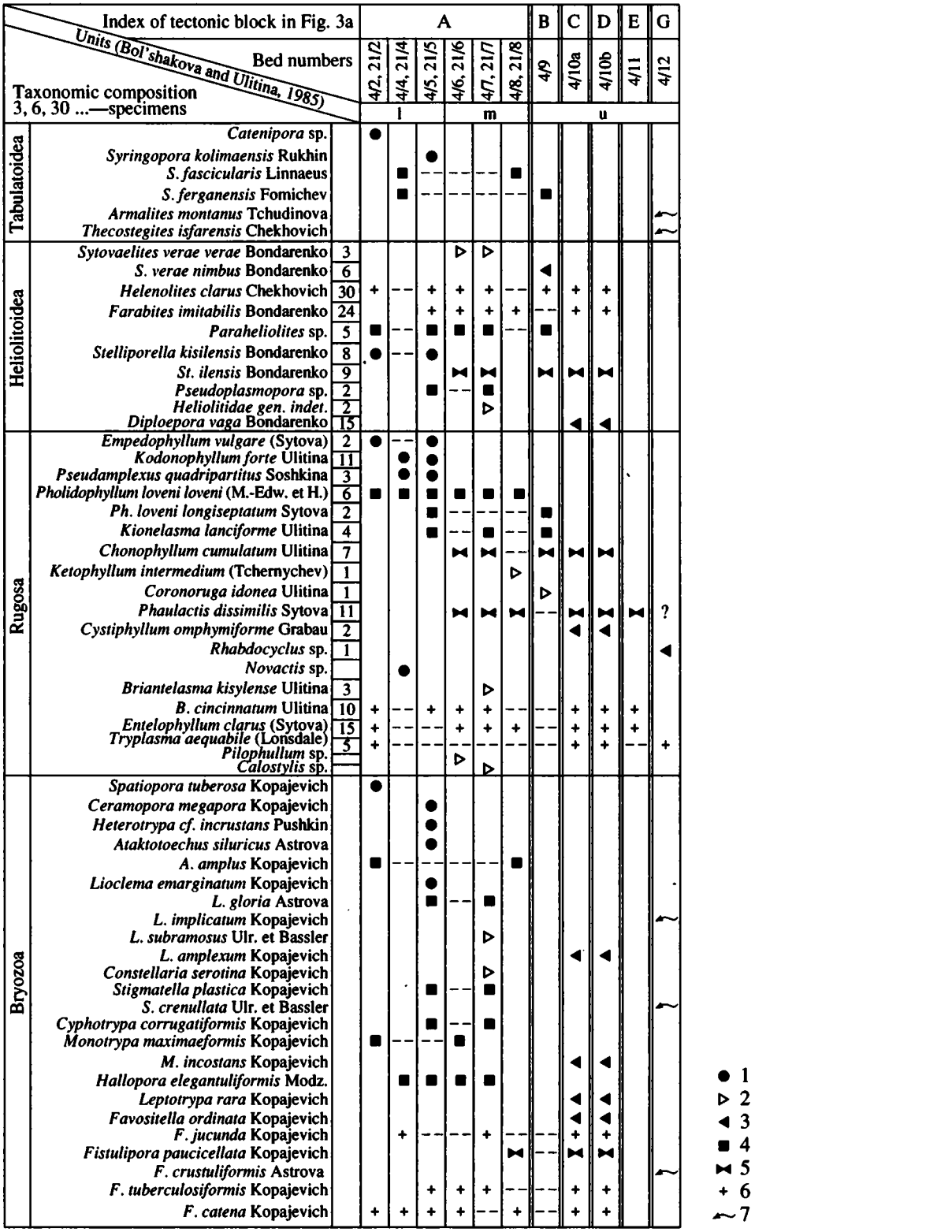

Fig. 6. Distribution of tabulates, heliolitids, rugoses, and bryozoans in the Chokusu Beds, tectonic blocks A–G of the Mt. Kyzyl-Dzhar-Chokusu (list of taxa after Sytova and Ulitina (1983); Kopaevich (1984); Bol'shakova and Ulitina (1985); Bondarenko (1986–1988, 1991, 1992, 1999); new data of Bondarenko are also included). Units of the Chokusu Beds: (l) lower, (m) middle, and (u) upper (subdivision after Kopaevich and Ulitina, 1977, and after Bol'shakova and Ulitina, 1985); species found in the lower (1), middle (2), upper (3), lower and middle (4), middle and upper (5), lower, middle and upper (6) units; (7) unspecified bed (figures in the second column indicate number of specimens).

This bed is a good marker for searching fauna during the field work.

Rugosa of Block IX have not been described. Ulitina distinguished three assemblages of rugoses in blocks A–D (Kopaevich and Ulitina, 1977; Sytova and Ulitina, 1983; Bol'shakova and Ulitina, 1985). Only *Empedophyllum vulgare* (Sytova), *Kodonophyllum forte* Ulitina, *Pseudamplexus quadripartitus* Soshkina, and *Novactis* sp. are components of the first assemblage (Fig. 6: beds 4/2–4/5 of Block A). Second and third assemblages include many other forms. Species *Chonophyllum cumulatum* Ulitina, *Ketophyllum intermedium* (Tchernychev), *Phaulactis dissimilis* Sytova, *Briantelasma kisylense* Ulitina, and *Calostylis* sp., all unknown in the first assemblage, appear in the second one (beds 4/6–4/8 of Block A). Tetracorals from Block B (Bed 4/9), where *Coronoruga idonea* Ulitina coexists with species occurring in the first, second and third assemblages, were distinguished by Ulitina as the second assemblage of a higher stratigraphic level (Bed 4/9) than that from beds 4/6–4/8. Ulitina regarded the third Rugosa assemblage (blocks C–G) as the youngest. In addition to species found in the first and second assemblages, it includes *Cystiphyllum omphymiforme* Grabau and *Rhabdocyclus* sp.

The analysis of *bryozoans* leads to the conclusions identical to those based on distribution of tetracorals (Fig. 6).

Rozman distinguished two *brachiopod* assemblages from deposits of blocks II and IV (Fig. 5). One of them, like in Block IX, is the atrypid–leptaenid assemblage (Bed 34c of Block II and Bed 29 of Block V), and the second was named the upper assemblage (beds 34a, b of Block II, beds 31a, b of Block IV, and Bed 27 of Block VII).

Thus, our analysis of fossil invertebrates from the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu and their stratigraphic relationships (older–younger) suggest the following inferences:

(1) Two successive assemblages of tabulates and brachiopods are distinguished in deposits of Block IX. The older assemblage consists of 13 species, and 17 taxa are identified in the second one. Nine forms are common for both assemblages.

(2) One assemblage of tabulates, two assemblages of heliolitids, three assemblages of tetracorals and bryozoans, and two assemblages of brachiopods can be recognized in deposits of blocks II–III and A–G. Some of them are likely concurrent, and their distinctions seem to be related to different habitat environments.

(3) Reappearance of assemblages in rocks of blocks II–VIII and A–G offers two interpretations. According to one of them the blocks are coeval. The second implies the recurrence of assemblages at different stratigraphic levels of the section, when the same habitat environments appear in the basin in the course of its evolution.

Age of the Chokusu Beds

Age of the Chokusu Beds estimated on the basis of different fossil groups leads to the contradictory conclusions (see Figs. 1 and 4). We see three reasons of this situation: (1) the beds bear few species in common with those known from the type sections and type localities of Silurian stages of the general stratigraphic scale; (2) many of the species are established so far in Mongolia only; (3) the correlation is based on regional scales, where stage subdivisions are not always adequately substantiated. Nevertheless, we should take into consideration first of all species in common with those known from England, the type region for almost all the Silurian stages (except the Pridolian). Another region of great importance for correlation is Sweden (Gotland Island) that was incorporated during the Silurian along with England into a single European paleogeographic province comprising Estonia and Podolian Dniester region as well. The Silurian species of the European province are also found in Mongolia that was a part of the Central Asian Province at that time.

Tabulates (see Figs. 5, 6). The early lists of tabulates from the Chokusu Beds appeared in the 1980s (Bol'shakova and Ulitina, 1985; Chudinova, 1986; Rozman and Minzhin, 1988).

The comprehensive study of tabulates allowed Minzhin to distinguish 16 species in the Chokusu Beds. The comparison between lists in Figs. 5 (16 species) and 4 (6 species) shows that the genus *Syringopora* is presented in both, though its species are different. Eight species of the genus are identified, one of which is new. Beyond Mongolia, age of these species is interpreted as follows (Chudinova, 1986): *S. kolimaensis* Rukhin—the Late Silurian–Early Devonian of northeastern Russia; *S. fasciculatis* Linnaeus—the Wenlockian–Ludlovian of England, Kuznetsk basin, Novaya and Severnaya Zemlya archipelagos, or the Ludlovian (Lockportian) of North America and Central Asia, and the Ludlovian–Pridolian of the Podolian Dniester region; *S. ferganensis* Fomichev—the Late Silurian of Central Asia (Fergana and the Isfara River, upstream of the Matcha Village); *S. gorskyi* Tchernychev—the Wenlockian of northeastern Russia (the Sandugan Horizon in the Kolyma and Yasachnaya river basins), or the Pridolian of the Podolian Dniester region and the Late Silurian of Novaya Zemlya and Central Asia; *S. khalagensis* Tchernychev—the late Llandoverian–Wenlockian of the Sette Daban Ridge, or the Late Silurian of Novaya Zemlya Archipelago and Vaigach Island; *S. scabra* Sokolov—the Wenlockian of the Siberian platform, or the Ludlovian of the Bol'shezemel'skaya Tundra; *S. novella* Klaamann—the Llandoverian–Wenlockian of Gotland (the Slite Beds), Estonia (the Yaani Horizon), and the Podolian Dniester region (the lowermost Kitaigorod Horizon), or the Wenlockian–Ludlovian of Tuva and the Wenlockian of northeastern Russia (the Anika and Sandugan horizons).

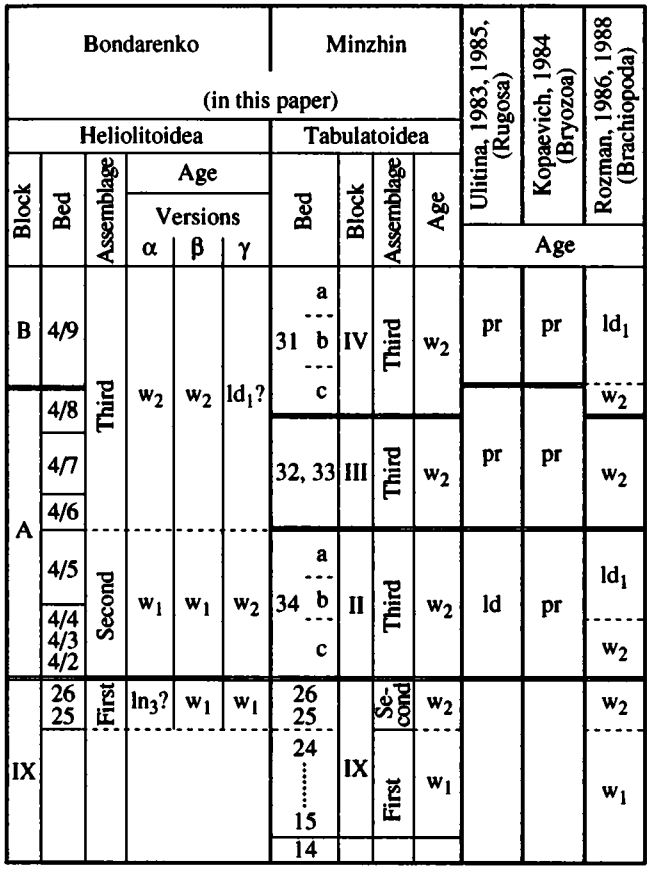

Fig. 7. Correlation of tectonic blocks with the Chokusu Beds and age interpretation of the latter on the basis of different groups of invertebrates.

In addition to the *Syringopora* genus, two more taxa from the order Siringoporida are mentioned in publications on the Chokusu Beds: "Encountered throughout the entire section are ... *Thecostegites isfarensis* Chekhovich and *Armalites montanus* Tchudinova" (Bol'shakova and Ulitina, 1985, p. 33). These genera and species have not been found in samples collected by Minzhin. The *Thecostegites* genus is characteristic of the Middle–Late Devonian; its occurrence in Ludlovian strata is reported from the Bol'shezemel'skaya Tundra and Central Asia. *T. isfarensis* Chekhovich was found in the upper Ludlovian Isfara Beds of the southern Fergana Valley. The genus *Armalites* (the Late Silurian–Middle Devonian, Eifelian) is represented by *A. montanus* Tchudinova from Ludlovian rocks of Tuva (Mt. Elegest). Thus, according to Chudinova, the taxonomic composition of the order Siringoporida indicates the Late Silurian age of the Chokusu Beds, however, siringoporids identified by Minzhin imply the Wenlockian age of the beds.

The presence of halysitids of the genus *Catenipora*, which are listed by Bol'shakova and Ulitina (1985), finds no support. Among two *Halysites* species identified in the Chokusu Beds, *H. senior* Klaamann is known from Wenlockian deposits of the Gotland Island and Estonia (upper beds of the Yaani Horizon) and *H. junior* Klaamann from the Wenlockian of Gotland, Estonia, and northeastern Russia.

Favositids are represented by five genera and nine species. The *Palaeofavosites asper* Orbigny species was found in the Wenlockian deposits of England, Gotland, Estonia, and Poland. *Palaeofavosites porosus* Sokolov is known from the Llandoverian deposits of Central Asia (Zeravshan Ridge), Tuva, northeastern Russia, as well as from the Wenlockian deposits of Estonia (the Yaani Horizon) and Kazakhstan (Tarbagatai Ridge). *Mesofavosites obliquus* Sokolov is characteristic of the Llandoverian deposits of Estonia (the lowermost Adavere Horizon), Taimyr Peninsula, Siberian platform, eastern Kazakhstan, and Altai Mountains, or of the Wenlockian deposits in Estonia and Podolian Dniester region; in the Sette Daban Ridge and northeastern Russia, this species occurs in Lower Silurian deposits. *Favosites antiquus* Sokolov is encountered in the Wenlockian deposits of Estonia (the Yaani Horizon), and *Favosites jaaniensis* Sokolov is found in the Llandoverian (?)–Wenlockian deposits of the Podolian Dniester region, in the Wenlockian rocks of Estonia, and in the Ludlovian strata of Kazakhstan. The *Tuvaelites hemisphaericus* (Tchernychev) is buried in the Ludlovian deposits of Tuva (the Dashtygoi Beds) and Podolian Dniester region (the Malinovtsy Horizon), or in the analogs of the Wenlockian deposits in Siberian platform and Vaigach Island (the Greben' Horizon). *Barrandeolites bowerbanki* (M. Edw. et H.) is distinguished in the Wenlockian deposits of England (Dudley locality) and Czechia (the Kopane Beds).

Thus, tabulates from the Chokusu Beds are of the Wenlockian age that is well confirmed by many species (11 of 16) identical to the Wenlockian forms of the European Province, where the type section of the stage is located.

In Mongolia, tabulates of the Chokusu Beds have much in common with their counterparts from the Wenlockian Barun-Urt Beds of eastern Mongolia. Minzhin refers the first assemblage of tabulates from the Chokusu Beds to the early Wenlockian, and the second and third assemblages to the late Wenlockian.

Heliolitids. Sharkova was first to identify heliolitids (*Heliolplasma* sp.) in the Mt. Kyzyl-Dzhar-Chokusu section (Dergunov *et al.*, 1980). Classical descriptions of heliolitids appeared since 1986 (Bondarenko, 1986, 1987, 1988, 1991, 1992, 1999). At present, nine heliolitid species from eight genera are recognized in the Chokusu Beds (Figs. 5, 6). Four of eight genera are cosmopolitan (*Stelliporella, Diploepora, Paraheliolites, Pseudoplasmopora*), another four genera were distinguished quite recently, and their spatial and time ranges are unknown as yet (*Helioplasmolites* Chekhovich, 1955; *Helenolites* Chekhovich, 1977; *Sytovaelites* Bondarenko, 1986; *Farabites* Ospanova, 1989). Dating the Chokusu Beds, we took into consideration first of

all the heliolitids known from the type locality. Second, their astogeny, phylogeny, and mutability were studied as a basis for distinguishing the phylogenetic stages (phylozones). Third, the correlation was based on these phylogenetic stages. Fourth, we analyzed the validity of age determinations for deposits in the regions, where recent stratigraphic schemes have been elaborated (Altai, Tuva, northeastern Russia, China).

The most remarkable among cosmopolitan genera are *Diploepora* and *Stelliporella* (Bondarenko, 1987, 1991, 1992; Hill, 1981). The oldest specimens of *Diploepora* are encountered in the Upper Ordovician (Ashgillian) deposits of Norway and in the Llandoverian strata of Tuva and Altai. The genus became prosperous in the Wenlockian, when it was one of the main reef-building genera especially in the European Province (England, Scotland, Sweden, Estonia). The genus became extinct in the early Ludlovian. It is unknown in the Ordovician of the Central Asian Province, whereas in the Silurian of the province, it is represented by *D. vaga* Bondarenko. The last species was found in the upper Alash and overlying beds of the upper Chergak Subhorizon of western Tuva (the middle Llandoverian–Wenlockian), in the Yarovka Horizon of the Altai Mountains (the upper Llandoverian), and in the Barun-Urt Beds of eastern Mongolia (the Wenlockian). In the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu, *D. vaga* species was recovered from biohermal limestones of beds 4/10a and 10b (blocks C and G). The evolutionary level of their colony corresponds to the Wenlockian phylogenetic stage in the species evolution. We should also point out that *D. vaga* was found in the type section of the Yarovka Horizon in the Altai Mountains, the late Llandoverian age of which is initially established on the basis of brachiopods (Kulkov, 1967) and then confirmed by distribution of graptolites (Sennikov, 1976) and conodonts (Vorozhbitov, 1996). Judging from the conodont assemblages, "the section likely spans the upper part of the *celloni* Zone ... and the lower part of the *amorphognathoides* Zone" (Vorozhbitov, 1996, p. 103). The evolutionary level of *D. vaga* colonies in the Yarovka Horizon of the Altai Mountains corresponds to a younger stage of the species evolution as compared to that of its predecessors from the Chokusu Beds of Mongolia (Bondarenko, 1987), and this confirms once again the younger age of the beds. In addition, brachiopod assemblages coexisting with *D. vaga* in the type section of the Yarovka Horizon of the Mt. Rossypnaya (*Pentamerus* and *Pentameroides* species) and in the type section of the Chokusu Beds (*Atrypa* and *Sulcatina* forms from blocks C and G) differ from one another.

The oldest species of the *Stelliporella* genus are established in the Upper Ordovician (the middle Ashgillian) strata, the youngest in the Ludlovian and, presumably, in the Pridolian. The prosperity time of the genus was in the Wenlockian, because its species are known from the following stratigraphic units: the upper Dudley reef limestone in England, the Barrandien Liten Formation in Czechia, the Marianovka Subhorizon of the Kitaigorod Horizon in the Podolian Dniester region, Beds "c" of Gotland, and also from the Wenlockian strata of many other regions. The genus became extinct in the Ludlovian. Its presence in the Pridolian deposits still remains to be proved. Two species of the genus are established in the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu (Bondarenko, 1991): *St. kisilensis* Bondarenko and *St. ilensis* Bondarenko. The latter was found in the Barun-Urt Beds of eastern Mongolia (the Wenlockian) and in Silurian deposits of Tuva (the stage is unknown). *St. kisilensis* belongs to the second heliolitids assemblage, and *St. ilensis* is a component in the second and third assemblages.

The genus *Helenolites* was identified by Chekhovich (1977) in the lower part of the Pichi-Shui Beds of Tuva. Four species of the genus are known. Three species corresponding to the first phylogenetic stage are *H. clarus* Chekhovich along with its younger synonym *H. mutabilis* Chekhovich, *H. foroensis* (Lindström), and *H. wentzeli* (Galle). *H. tuvellus* Chekhovich corresponds to the second stage of phylogeny. The most confident ages are known for *H. wentzeli* (Galle) and *H. foroensis* (Lindström). The former was found in the upper part of the Barrandien Liten Formation of Czechia (the Wenlockian), and the latter is described from the Slite and Klintenberg Beds of Gotland and Förön islands, Sweden (the middle Wenlockian–lower Ludlovian). *H. clarus* Chekhovich is of the same evolutionary level as *H. wentzeli* (Galle) and *H. foroensis* (Lindström), and the time range of the species is interpreted as the Wenlockian–early Ludlovian. As mentioned above, *H. clarus* Chekhovich is characteristic the Barun-Urt Beds in eastern Mongolia (the Wenlockian), and also of the lower part of the Pichi-Shui Beds in Tuva (the lower Ludlovian). In the Chokusu Beds, it is a component of all four assemblages and coexists with the genus *Helioplasmolites* in the first assemblage, with *Stelliporella* forms in the second and third assemblages, and with *Diploepora* in the fourth. Such a combination of genera is characteristic of the European Province as well, especially of the Wenlockian deposits in England, Sweden, and Czechia. In the succession, the combination of *Helenolites* and *Helioplasmolites* is older than that of *Helenolites, Stelliporella, Helenolites*, and *Diploepora*. Genera *Diploepora* and *Stelliporella* are missing in the lower part of the Pichi-Shui Beds of Tuva, the type locality of *H. clarus* Chekhovich. This may indicate a higher stratigraphic level of the species in Tuva, as compared to that in the Chokusu and Barun-Urt Beds of Mongolia. Most likely, the lower part of the Pichi-Shui Beds corresponds to the lower Ludlovian and, probably, to a part of the uppermost Wenlockian.

In general, heliolitids from the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu are of the Wenlockian age, but it is also probable that the first heliolitid assemblage corresponds to the late Llandoverian. Heliolitids sug-

gest the closest similarity of the Chokusu Beds to the Wenlockian Barun-Urt Beds of eastern Mongolia.

Rugosa. Ulitina (Kopaevich and Ulitina, 1977; Sytova and Ulitina, 1983; Bol'shakova and Ulitina, 1985) considers rugose corals from the Chokusu Beds of blocks A-G as the Ludlovian–Pridolian in age. It should be pointed out that, when correlated, Rugosa of the Chokusu Beds are similar to the late Llandoverian, Wenlockian (first of all), and earliest Ludlovian tetracorals of the European Province, especially to those from England and Sweden (Gotland). The situation changes in the case of correlation with tetracorals from Tuva. Among the rugose forms from the Chokusu Beds, there are some taxa known as well from the Pichi-Shui Beds of Tuva that are attributed to the Pridolian by some researchers and to the Ludlovian by others.

Bryozoans. The collection of bryozoans from the Chokusu Beds is even more contradictory than that of rugose corals (Kopaevich and Ulitina, 1977; Kopaevich, 1984). Seventeen of twenty one bryozoan species are described as new, one is classed in the open nomenclature, one more is known from the Upper Ordovician, two others from the Upper Silurian, one from the Upper Silurian–Lower Devonian, two from the Lower Devonian. Kopaevich (1977) originally assigned the bryozoan assemblage of the Chokusu beds to the Pridolian (?)–Gedinnian, and later (1984) to the Pridolian.

Brachiopods. On the basis of this faunal group, Rozman (1986, 1988) interpreted the age of the Chokusu Beds as the Wenlockian–early Ludlovian attributing the lower (gipidulid) assemblage to the late Wenlockian, and the upper one to the early Ludlovian. Rozman pointed out that the lower assemblage includes the late Llandoverian elements, whereas the upper is similar to that of the Pichi-Shui Beds (Kul'kov *et al.*, 1985). It should be also noted that the late Llandoverian– early Wenlockian (?) brachiopods coexist in southern Mongolia with conodonts of the *celloni* and *amorphognathoides* zones (Vorozhbitov, 1997). The southern assemblage of brachiopods differs from that of the Chokusu Beds of northwestern Mongolia: only four of twenty genera (*Isorthis, Leptaena, Stegerhynchus, Tannuspirifer*) and three of thirty eight species (*Stegerhynchus angaciensis* Tchernychev, *St.* ex gr., cf. *St. borealis* Sepl., *Tannuspirifer pedaschenkoi* Tchernychev) are common for both regions.

CONCLUSION

(1) Fossil assemblages of the Chokusu Beds are closely related to each other and have many species in common. Their interrelations suggest a short time range of development, similar habitats (a shelf zone), and the lack of appreciable hiatuses. Differences in composition of fossil assemblages imply a successive change of faunas and sedimentation environments in the history of the basin area under discussion, and also faunal and facies changes along the strike in the shelf zone.

(2) When compared with taxa of the European Province, tabulates, heliolitids, rugose corals, and brachiopods from the beds reveal similarity to their assemblages from the upper Llandoverian, Wenlockian, and upper Ludlovian deposits, especially from the Wenlockian strata. When correlated with faunas from Tuva, the studied assemblages reveal species in common with the Pichi-Shui Beds attributed to the Ludlovian–earliest Pridolian on the basis of brachiopods, stromatoporoids, corals, bryozoans, and ichthyofauna.

(3) According to combination of genera and evolutionary stages, heliolitids from the Pichi-Shui Beds of Tuva correspond to a higher stratigraphic level than that of the Chokusu Beds of the Mt. Kyzyl-Dzhar-Chokusu.

(4) Heliolitids from the Chokusu Beds are most similar to this fauna from the Barun-Urt Beds of eastern Mongolia, where the latter are dated back to the Wenlockian on the basis of different groups of invertebrates.

(5) We consider the Chokusu Beds as Wenlockian in age but the early Llandoverian age of their lower part and the early Ludlovian age of the upper one are also possible (Fig. 7).

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The Reference Section of Silurian–Lower Devonian Boundary Deposits in the Eastern Slope of the Urals

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Abstract—Carbonate and volcanogenic–sedimentary deposits composing the upper Pridolian Substage and Lochkovian stages in the area between Boksity Station in the south to the Sos'va River in the north (eastern slope of the northern Urals in the vicinity of Severoural'sk) are described with consideration of ranges, boundaries, distribution, and lithological and paleontological characteristics of the Severoural'sk, Sarainaya and Sauma horizons. Correlation of carbonate and volcanogenic–sedimentary successions, and formation history of the whole sequence are also discussed.

Key words: Pridolian, Lower Devonian, Severoural'sk, Sarainaya, and Sauma horizons, section.

INTRODUCTION

The article is written at the suggestion and wish of our friend and colleague Mariya Vasil'evna Shurygina after her death and is based on the materials considered in the unpublished report "The Silurian–Lower Devonian reference section of the Petropavlovsk bauxite-bearing structural–facies zone in the eastern slope of the Urals" (1982). Shurygina was principal executor of the report and author of the sections "Stratigraphy" and "Section descriptions."

The Silurian–Lower Devonian section in the area of Severoural'sk Town (Fig. 1) is a unique in respect of the completeness and continuity of sedimentary succession, and of facies relationships between carbonate, volcano-sedimentary, and volcanogenic rock varieties.

The study of the reference section in the area of Severoural'sk Town was carried out in 1980–1982 by geologists from the paleontological–stratigraphic team of the Uralian geological–mapping expedition. The report included a detailed description of the reference section along the Kolonga and Vagran rivers and other sections exposed along Shegul'tan and Sos'va rivers, and also in ditches and quarries. Core sections of 11 deep wells were also studied, and the available abundant materials obtained earlier were analyzed. Important was a participation in the work of D.K. Patrunov, an expert in lithology from the All-Russian Research Institute of Geology and Mineral Resources of the World Ocean (St. Petersburg), who performed lithological study in the field and laboratory.

We used also results of paleontological studies of different fossil groups provided by the team of experts in paleontology and stratigraphy: F.E. Yanet and L.I. Loburtseva (tabulates), M.V. Shurygina (tetracorals), I.A. and M.G. Breivel' (brachiopods), N.Ya. Antsygina (trilobites), G.G. Zenkova (ostracods), and V.S. Militsina (crinoids). Stromatoporoids were determined by O.V. Bogoyavlenskaya from the Uralian State Mining–Geological Academy, algae by V.P. Shuiskii from the Institute of Geology and Geochemistry of the Uralian Division RAS and also by L.G. Petrova from our paleontological–stratigraphic team. Drawing is performed by I.S. Maksimova.

† Deceased.

GEOLOGICAL REVIEW

Paleozoic sediments widespread in the vicinity of Town of Severoural'sk received a keen attention, since the 1930s after discovery of their bauxite deposits. The history of geological investigations in the area is considered in detail in many works (Peive, 1947; Streis, 1951; *Biostratigrafiya i fauna...*, 1977; Shurygina *et al.*, 1981; and others).

This paper considers deposits of the Petropavlovsk Formation and its age analogues. The Petropavlovsk Formation was first defined by C.M. Andronov (unpublished reports), and later, with the reference to Andronov's data, by Peive (1947). Its description and characteristics of the Sos'va and Sarainaya formations are given by Peive (1947) and Andronov (1961).

As a result of study by the paleontological–stratigraphic team, the Petropavlovsk Formation was subdivided into three biostratigraphic zones (Breivel' *et al.*, 1971) and horizons. The distinguished Severoural'sk Horizon corresponds to the Silurian (Pridolian Stage), whereas the Sarainaya, and Sauma horizons are attributed to the Lochkovian Stage of the Devonian (Breivel' *et al.*, 1974; *Biostratigrafiya i fauna...*, 1977). In these and other works (Dubatolov *et al.*, 1968; Shurygina

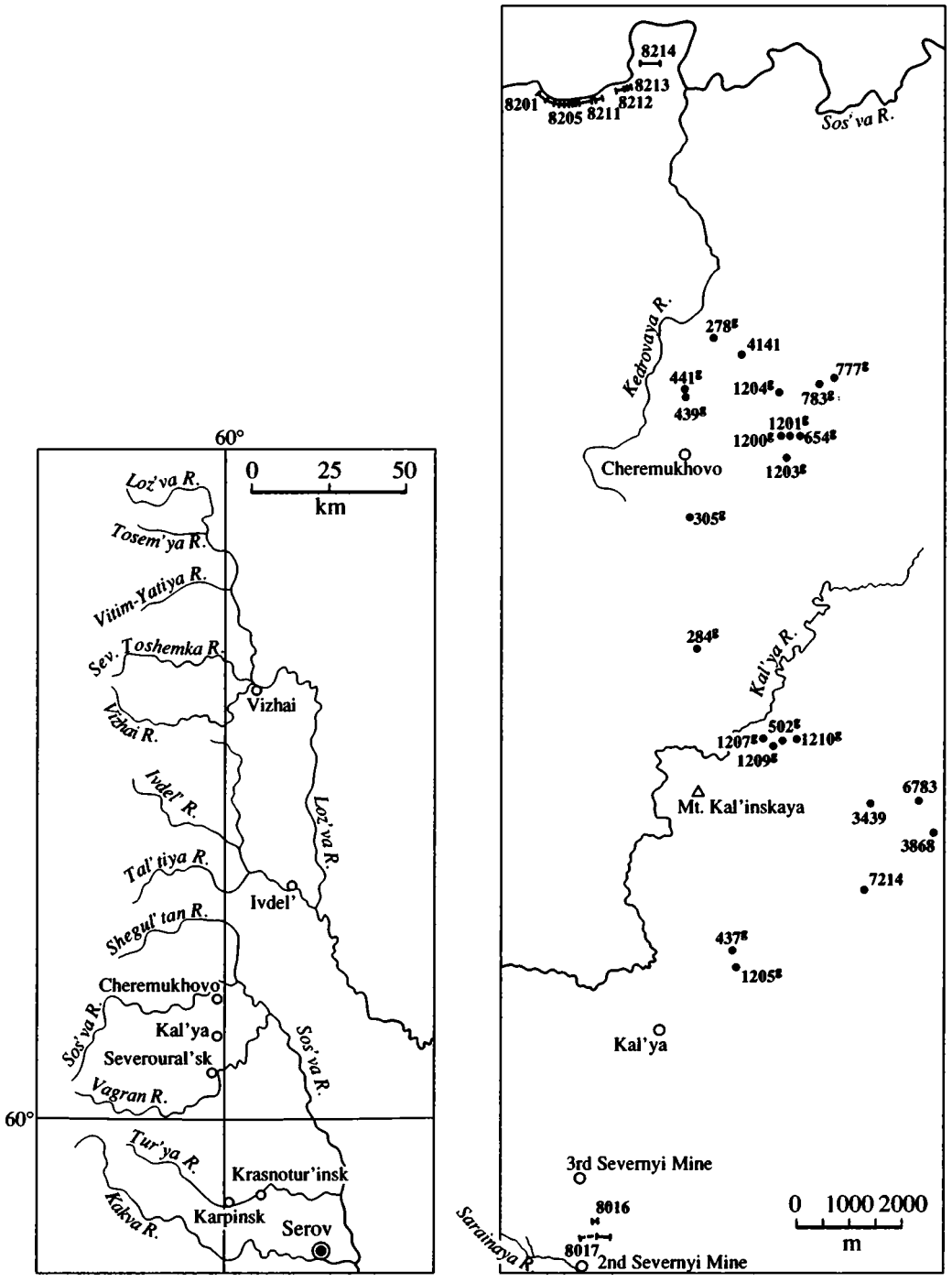

Fig. 1. Sketch map of the study area and location of wells and exposures to the north of Severoural'sk.

et al., 1981; and others), position of the Silurian–Devonian boundary in the eastern slope of the Urals, the Severoural'sk area included, and some problems of the interregional correlation were considered. The Silurian–Devonian boundary was placed at a lower level, because previous researchers referred to the Devonian only the uppermost part of the Petropavlovsk Formation underlying the bauxite-bearing unit. The remaining part of the formation was considered as a unit of the Silurian (Pronin and Khodalevich, 1947; Streis, 1951; and others).

SECTION DESCRIPTION

Severoural'sk Horizon

The Severoural'sk Horizon (Breivel' *et al.*, 1976; *Biostratigrafiya i fauna...*, 1977) corresponds to the

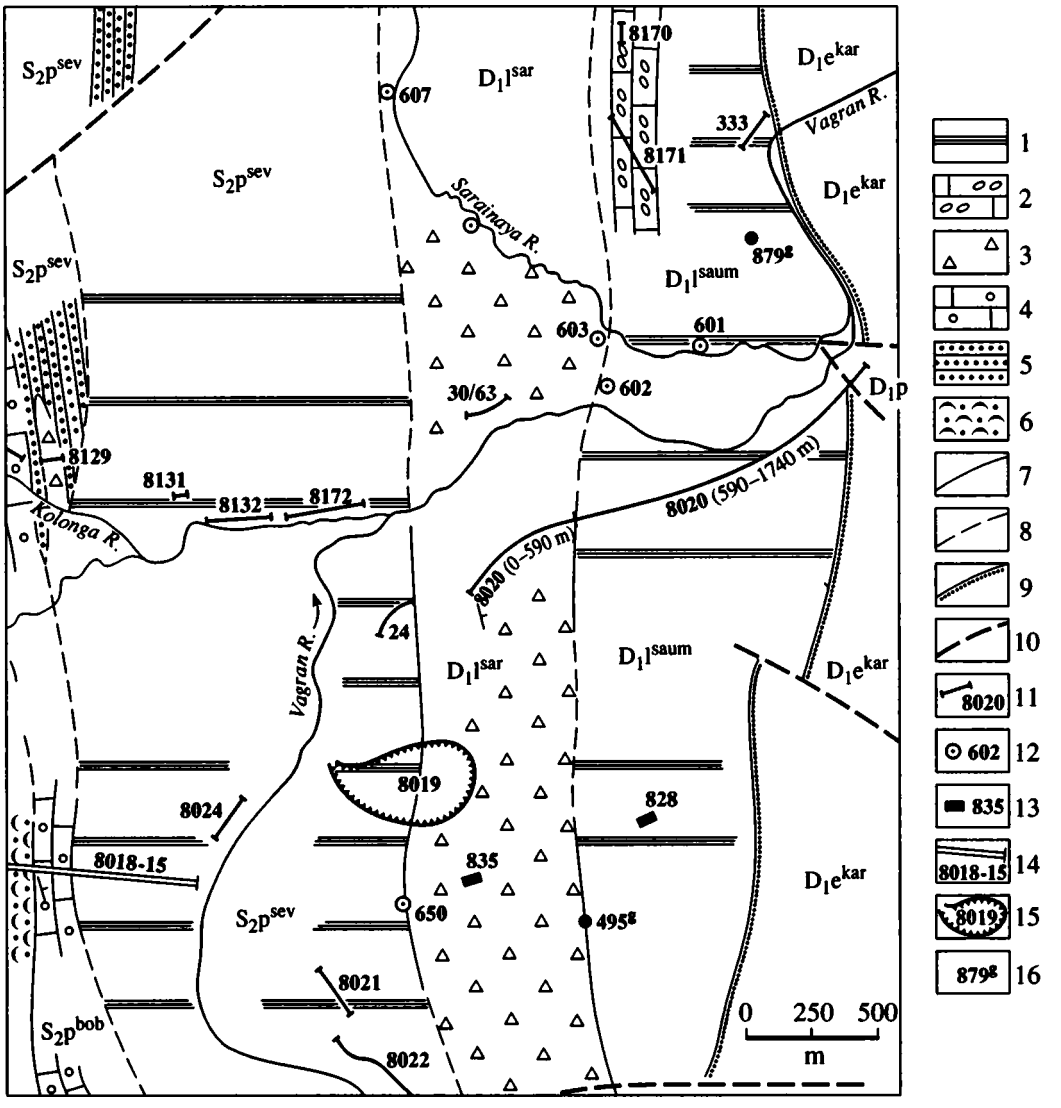

Fig. 2. Schematic geological map of the Severoural'sk area: (1) pale massive limestone, mainly algal; (2) bedded cloddy algal limestone; (3) massive carbonate breccia; (4) granular skeletal–detrital limestone; (5) volcanomictic sandstone; (6) tuffstone and tuffaceous conglomerate; (7–9) stratigraphic boundaries, (7) proven, (8) assumed, and (9) unconformable; (10) tectonic boundaries; (11, 12) exposures, (11) in scale and (12) out of scale; (13) prospecting shaft; (14) ditch; (15) quarry; (16) well; (S_2p^{bob}) Upper Silurian, Pridolian; Bobrov Horizon; (S_2p^{sev}) Upper Silurian, Pridolian; Severoural'sk Horizon; (D_1l^{sar}) Lower Devonian, Lochkovian, Sarainaya Horizon; (D_1l^{saum}) Lower Devonian, Lochkovian, Sauma Horizon; (D_1p) Lower Devonian, Pragian; (D_1e^{kar}) Lower Devonian, Emsian, Karpinsk Horizon.

lower part of the Petropavlovsk Formation exposed along the Kolonga and Vagran rivers, where the horizon stratotype is located (Shurygina *et al.*, 1981). The lower part of the horizon is recovered by the drainage trench in right-hand side of the Istok River (exposure 8018, Bed 15), and its upper part is exposed in a large quarry in the right-hand side of the Vagran River (exposure 8019). The middle portion of the horizon section is buried under the Vagran River valley. This unexposed interval corresponds to limestone exposures along the left bank of the Kolonga River near the church (exposures 8132 and 8172, see Fig. 2).

The Severoural'sk Horizon is variable in lithological aspect. Along the Kolonga River, it is represented by a sequence of light-gray massive limestone beds 470–500 m thick in total. Northward and southward, between the Kolonga and Sos'va rivers, limestone beds are replaced by dark-colored tuffstone and tuffaceous siltstone of the Sos'va Formation 160–170 m thick (Well 1205g). Synchronism of tuffaceous rocks and limestone sequence is substantiated by their stratigraphic position between limestones of the Bobrovka Horizon and tuffaceous conglomerates with the fauna of the Sarainaya Horizon. The thickness of carbonate

rocks exposed along the Sos'va River is about 100 m, but, the horizon includes here also the volcano-sedimentary strata. The carbonate section of the horizon exposed along the right-hand side of the Shegul'tan River is close to the stratotype in the thickness.

The limestone succession of the Severoural'sk Horizon is underlain by bedded limestones of the Bobrovka Horizon also referred to the Pridolian Stage. In the upper part of the Bobrovka Horizon that is exposed in the left-hand side of the Vagran River, downstream of the pumping station and southward of Severoural'sk, the medium- (30–40 cm) to thin-bedded (7–15 cm) limestones (exposure 8004) grade upward in the section into thick-bedded (50–60 to 80–100 cm) limestones colored lighter (interval 2) and overlain by white massive limestones of the Severoural'sk Horizon (interval 3). In the stratotype section (exposure 8018, Bed 15), the bedded limestones of the Bobrovka Horizon occur below the highly disintegrated silty–calcareous rock (0.8–1 m) and thick-bedded to massive limestones. About 1.5 km to the north, in the left-hand bank of the Kolonga River, the basal part of the massive limestone unit is represented by a thick (to 20 m) member of tuffstone and tuffaceous siltstone with interbeds of carbonate breccia composed of small rock fragments (exposure 8129). The boundary is marked by faunal changes: the Severoural'sk Horizon bears rare stromatoporoids, solitary Rugosa corals are replaced here by colonial forms, brachiopods and crinoids are more diverse, and ostracods changed their composition and became less abundant.

Northward, between the Kolonga River and the Kal'ya Settlement (Figs 2, 3), the lower boundary of the Petropavlovsk Formation probably corresponds to the base of the Sos'va Formation. Sections in this area show gradual transition from the upper limestone member of the Bobrovka Horizon to tuffstones and tuffaceous siltstones. The contact was recovered by mining drifts near the Kolonga River in the vicinity of the 2nd Severnyi Mine and Kal'ya Settlement (Well 1205^{g}).

In the area between the Bol'shaya Kal'ya and Sos'va rivers, rocks of the Bobrovka Horizon are replaced by the Sos'va Formation composed of volcanogenic rocks variable in lithology. The lower boundary of the Sos'va Formation in these sections does not coincides with the top of the Bobrovka Horizon and is located lower, thus corresponding to the top of the lower member of the horizon. Limestone interbeds from the lower part of the Sos'va Formation (exposures along the Sos'va River downstream of the Mostovaya River mouth) yielded corals (*Squameofavosites thetidis* Chekh. and *Ketophyllum amplexoidum* (Tschern.)) and coquina accumulations of *Atrypella camelina* (Buch.) correlating the rocks with the Bobrovka Horizon (Shurygina *et al.*, 1981). The position of the lower boundary of the Severoural'sk Horizon in these section is unclear, because the contact between the member of light massive limestone of the Severoural'sk Horizon and underlying strata, which are exposed along the Sos'va River nearby the Sos'va Settlement, is tectonic.

The lower portion of the Severoural'sk Horizon exposed along the Vagran and Kolonga rivers (exposure 8018, Bed 15, interval of 0–380 m) is composed of massive limestone, pale to white and rich in *Fistulella* remains with visible problematic tubular structures, and also of algal, less commonly, cloddy limestone with separate skeletal fragments, which show borer traces of Cyanophyceae. In addition, fragments of siphon algae and local accumulation of oval coprolites are abundant among microtextural elements.

The base of the horizon is marked by variably recrystallized fine-grained lithoclastic–skeletal–detrital limestones with algal matrix. The skeletal material is mainly of the crinoid origin. Crinoid limestone contains thin interbeds of thin-platy rocks with abundant spiny spicules—spongolite (Well 879^{g}, depth interval of 1558–1564 m). Common are encrustations of several generations.

The upper and thickest (about 300 m) portion of the horizon is characterized by the higher share of granular skeletal–detrital rock varieties, although the main rock types are the *Fistulella*, algal, and, to a lesser extent, cloddy limestones. Granular varieties with crinoid detritus locally grade into calcareous sandstone. Peculiar of this interval are widespread limestone breccias, abundant and variable encrustation structures, and local irregular bedding (40–50 cm to 1 m) with uneven, wavy bedding planes. Breccias are coarse, with small-sized rock fragments; the latter are angular, unsorted, and composed of algal limestone. The matrix is intensely pigmented by Fe hydroxides. The size of rock fragments ranges from several to 20–30 cm and more. Volcano-sedimentary rocks of the Severoural'sk Horizon are exposed from the 2nd Severnyi Mine to the Sos'va River. In the Kal'ya and Novaya Kal'ya areas, they were penetrated by wells 1205*g* and 502*g*, which recovered the relatively uniform volcanogenic–terrigenous sandstones, gravelly sandstones, siltstones, and shales intercalated with subordinate thin gravelstone interbeds. The characteristic feature of these rocks is their rhythmical, frequently very thin bedding. A slightly different section is exposed along the Sos'va River (exposure 8201), where it is characterized by the coarse banding patterns in different lithological facies and by the coarse-grained texture (gravelstones intercalated with subordinate thin sandstone interbeds). In the upper part of the horizon, there are basaltic lava flows (Well 502^{g}, depth interval of 1633–1639 m).

Main rock-forming organisms of algal limestones from the Severoural'sk Horizon are tubular coral forms *Fistulella variabilis* Shuysk. and also Cyanophyceae: *Ikella vermicularis* Shuysk., *Girvanella maslovi* Shuysk., *G. ducii* Wether, *G. conferta* Champ., *Renaclis devonicus* Johns., *Wetheredella silurica* Wood., and *Ludlovia multispora* Korde. In total, more than 20 spe-

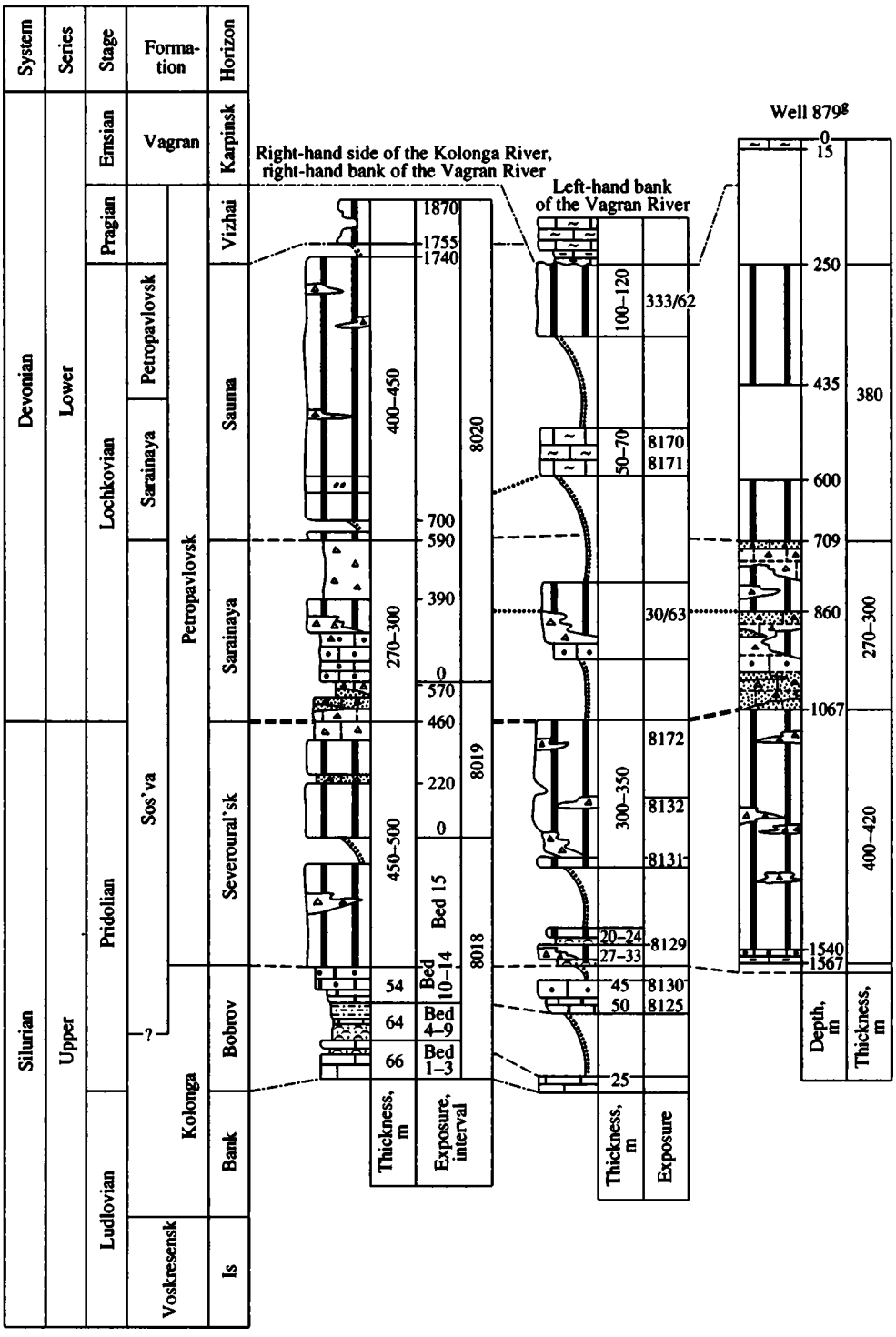

Fig. 3. Correlation of sections: (1) pale limestone, massive, generally algal; (2) gray to light gray, cloddy limestone, massive or with obscure bedding; (3) bedded and cloddy algal limestone; (4) carbonate breccia, (a) massive and (b) bedded; (5) limestone, (a) massive and (b) coarse-bedded biomorphic–detrital (including crinoid detritus and lithoclasts); (6) limestone, (a) granular, organogenic–detrital, with obscure or coarse bedding, and (b)carbonate breccia; (7) carbonate breccia with sandy matrix; (8) calcareous sandstone variably pigmented by Fe-hydroxides; (9) limestone, granular, skeletal–detrital; (10) skeletal–detrital limestone with muddy matrix; (11) skeletal–detrital–muddy limestone; (12) limestone, thin-bedded, cloddy and substantially muddy; (13) silty–clayey limestone or marl; (14) *Amphipora* limestone; (15) silty limestone or calcareous sandstone; (16) tuffaceous conglomerate; (17) tuffstones with gravel and pebbles; (18) tuffstone; (19) volcanomictic sandstone; (20) clayey–silty–sandy rocks; (21) silty–clayey rocks; (22) flyschoid member of alternating tuffstone, argillite, and siltstone; (23) volcanic breccia and large-block tuff of basaltic andesite composition; (24) basalt; (25) basaltic andesite; (26) andesite; (27) Silurian–Devonian boundary (a) and stage boundaries (b); (28) boundaries between horizons (a) and other correlation lines (b).

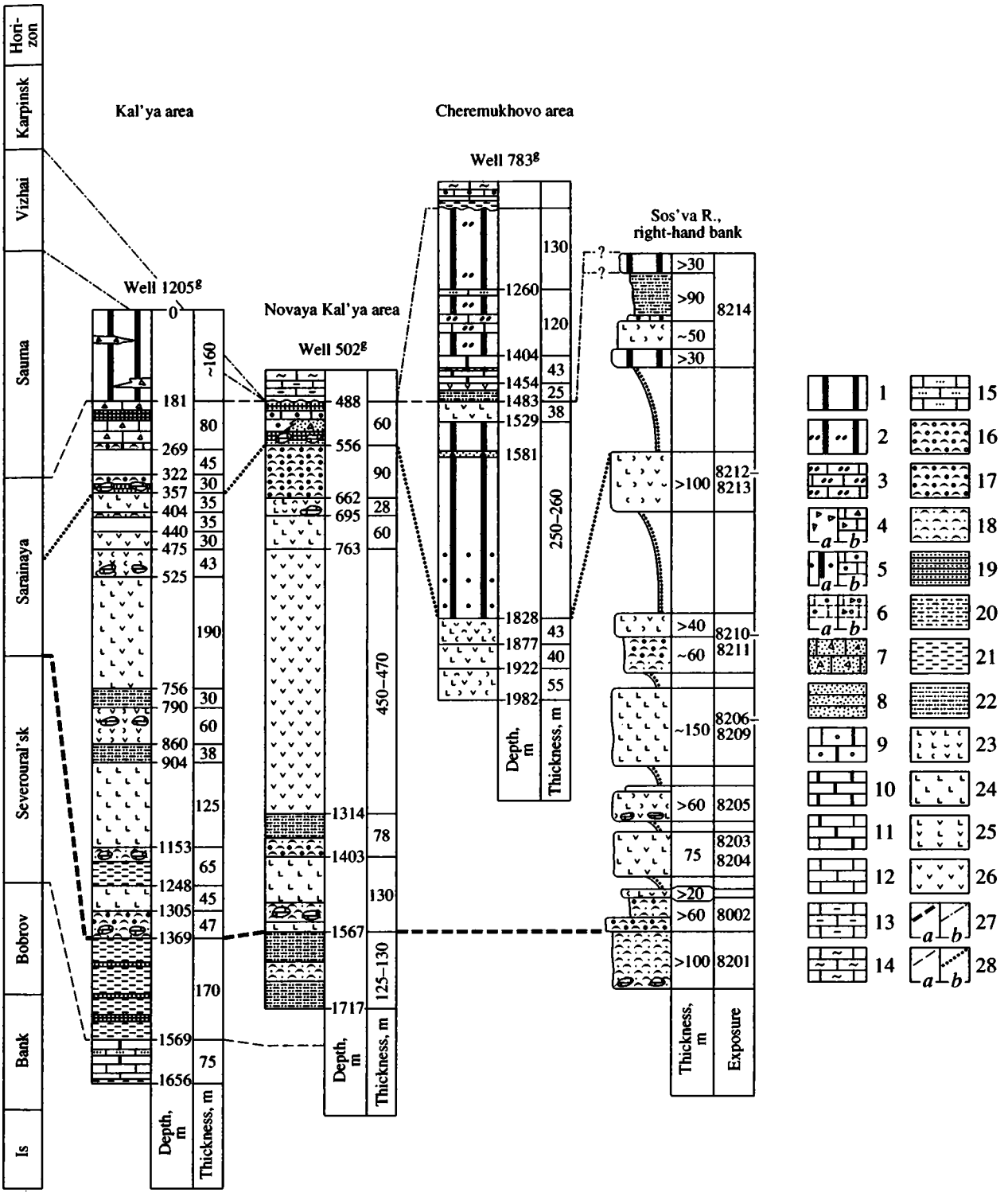

Fig. 3. (Contd.)

cies of Cyanophyceae, green and red algae, as well as algae of unclear taxonomy were determined.

The benthic fauna is distributed very irregularly, being confined mainly to detrital and organogenic–detrital limestones. It is particularly rare in the lower part of the horizon, where preponderant are algal limestones (exposure 8004–3; exposure 8018, Bed 15, interval of 0–380 m; exposure 8129; Well 879*g*, depth interval of 1400–1567 m). These layers yield rare iso-

lated coral colonies of *Favosites finitimus* Yanet, *F. bonus* Yanet, and *Pseudoamplexus* ex gr. *fascicularis* Soshk., and also brachiopods that are confined to the skeletal–detrital limestone varieties at the base of the horizon. Identified among brachiopods are species *Stropheodonta colongensis* Sap., *Levigatella olga* (Khod.), *Gypidula* cf. *uralica* Khod.), *G.* cf. *optata* Barr., *Uralotoechia kuschvensis* (Tschern.), *Atrypella cuboidoformis* (Khod.), *A. uraliaca* (Khod.), *A.* cf. *vagranica* (Khod.), *Lissatrypa columbella* (Barr.), *Spirigerina* ex gr. *marginalis* (Dalm.), and *Proreticularia pentameriformis* (Tschern.). Crinoids *Pisocrinius(?) costatus* (Schew.), *P.(?) astericus* (Schew.), *P.(?)* ex gr. *salebrosus* (Stuk.), *Syndetocrinus* sp., *Eucalyptocrinites(?) compositus* Milicina, *Mediocrinus medius* Stuk. and ostracods *Gavellina idonea* Abush., *Pseudorayella scala* Neck., *Tubulibairdia accurata* Zenk., *Turiella depressa* Zenk., and *T. minuta* Zenk. were collected at the same level.

In the upper part of the horizon (exposure 8018, Bed 15, interval of 380–420 m; exposures 8024 and 8019, interval of 0.0–460 m; exposures 8020-8023; Well 879*g*, depth interval of 1067–1400 m), the benthic assemblage is more abundant and diverse. It can be termed as the coral–brachiopod–crinoid assemblage. No thick populations of the coral meadow or biostrome type are noted, although sometimes corals are very abundant. Present are also small accumulations of brachiopods representing mainly genera *Lissatrypa, Tectatrypa, Atrypella*, and *Cyrtia*. Fragments of crinoid stems sometimes represent rock-forming components. Stromatoporoids and ostracods are fairly rare, and single trilobite specimens occur in addition.

The mentioned stratigraphic interval yielded the following groups: tabulates *Favosites finitimus* Yanet (abundant) and *F. bonus* Yanet, *Squameofavosites russanovi* (Tschern.); tetracorals *Circophyllum samsugnensis* Smith et Tremb., *Pseudoamplexus fascicularis* Soshk., *Astrictophyllum patrunovi* Shur. (abundant), *Entelophyllum* sp., *Rhabdacanthia crassiseptata* Shur., and *Holmophyllum subtenuis* Shur.; brachiopods *Gypidula uralica* Khod., *G. integer* Barr., *G. parvula* Tschern., *Levigatella olga* (Khod.), *Uralotoechia kuschvensis* (Tschern.), *Atrypella linguifera* (Khod.), *Lissatripa columbella* (Barr.), *Tectatrypa* cf. *tectiformis* (Tschern.), *Spirigerina* aff. *marginalis* (Dalm.), *Proreticularia pentameriformis* (Tschern.), and *?Cyrtia trapezoidalis* His.; ostracods *Gavellina idonea* Abush., *Saumella angusta* Zenk., *Tubulibairdia parva* Zenk., *Turiella elegans* Zenk., and *T. minuta* Zenk.; trilobites *Octobronteus khodalevichi* Weber; crinoids *Pisocrinus(?) astericus* (Shew.), *P.(?) costatus* (Schew.), *Parapisocrinus* ex gr. *ollula* (Angelin), *Crotalocrinites* sp. indet., *Syndetocrinus natus* Stuk. (fragments of arms, cups, and stems), *Eucalyptrocrinites(?)* cf. *subligatus* Milicina, *Desmidocrinus macrodactilus* Angelin, *Mediocrinus medius* Stuk., and *Pandocrinus* sp.

Limestones exposed along the Sos'va River in the vicinity of the Sos'va Settlement (exposure 8013) and along the Shegul'tan River are probably correlative with the upper part of the Severoural'sk Horizon.

Sarainaya Horizon

The stratotype section of the horizon is exposed along the Vagran River, from the outcrops located 500 m downstream of the old (natural) channel of the Kolonga River to the exposure situated 600 m further downstream (Breivel' *et al.*, 1976; *Biostratigrafiya i fauna...*, 1977). During the recent construction of water-protection system, the right-hand bank of the Vagran River was cleaned from loose deposits, and the section is now completely exposed. The base of the Sarainaya Horizon and its contact with the Severoural'sk Horizon is seen in a large quarry located in the right-hand side of the Vagran River upstream of the Kolonga River (exposure 8019, interval of 460–570 m); it was also penetrated by Well 879*g*.

The Sarainaya Horizon can be traced across the entire Petropavlovsk carbonate massif. The thickness of carbonate rocks in the stratotype section the horizon and in the deep Well 879*g* is up 270–300 m. In the vicinity of the 2nd Severnyi Mine, limestones of the horizon are about 100 m thick. The lower part of the horizon is entirely composed of tuffaceous conglomerates and tuffstones 500–550 m thick. In northern areas, the terrigenous–carbonate and volcanogenic members are thicker: the limestone thickness is only 80 m in Well 1205*g* near the Kal'ya Settlement and 30 m in the Mt. Kal'inskaya area. In Well 1205*g*, the volcanogenic and volcano-sedimentary rock succession includes rare limestone interbeds and is about 900 m thick. Easterly, in Well 783*g* located near the Cheremukhovo Settlement, the thickness of limestone beds is about 250 m, and, to the west of the settlement, it is approximately 50 m.

In the Petropavlovsk carbonate massif, the lower boundary of the horizon coincides with the level of transition from mainly algal limestones to carbonate breccia, calcareous sandstones, and organogenic–detrital limestones typical of the horizon. The limestone section of the Sarainaya Horizon encloses separate brecciated zones, which become dominant beginning from the depth of 380 m in Well 8019. The matrix pigmentation by Fe hydroxides is here more distinct, and encrustation structures are smaller. The faunal assemblage also changes owing to appearance of characteristic Devonian genera and species. The depth level of 460 m is marked by the first occurrence of Early Devonian forms *Favosites kozlowskii* (Sok.) and others. Beginning from the depth of 480 m, fossils occur more frequently, and at the depth of 570 m, they become abundant. The share of corals and crinoids increases, and colonial tetracorals are frequent rock-forming components. Pachyporids appear among tabulates. Crinoids are represented by diverse pisocrinids accompanied by first hexacrinids. The abundance of ostracods

also increases, particularly in sandy rock varieties. The depth level of 460 m is accepted for the lower boundary of the Devonian, although sedimentation patterns change at the depth of 380 m.

Northward, the Silurian–Devonian boundary is placed inside the Sos'va Formation and probably coincides with the level, where tuffstone beds are replaced by rudaceous tuffaceous conglomerates enclosing lenses and fragments of limestone with fauna characteristic of the Sarainaya Horizon (the depth interval of 1364–1365 m in Well 1205*g*, the Kal'ya Settlement area, and the depth interval of 1535–1538 m in Well 502^{g}, the Novaya Kal'ya Settlement area).

Rocks of the horizon show the well-developed clastic textures. Skeletal and lithoclastic rock varieties are represented by organogenic–detrital limestone and calcareous sandstones grading into carbonate breccia. Carbonate breccia is composed of large and small fragments of white algal and coral limestones. Fragments are to 30–35 cm across, unsorted, and angular. The matrix is calcareous, variegated. Breccia beds are especially characteristic of the basal part of the horizon, where their fine-grained sandy matrix is pigmented. Lentils (10–20 cm thick) of calcareous sandstones closely associate with breccia beds. Sandstones are composed of algal limestone lithoclasts and skeletal material. Grains are ferruginous, and detrital material is cemented by the algal and, sometimes, by micritic mass. Beds of loose and poorly cemented sands are present.

North of the Kolonga River, carbonate rocks slightly change. In the 3rd Severnyi Mine area, the upper part of the horizon is composed of algal limestone and carbonate breccia. In the vicinity of the Kal'ya and Novaya Kal'ya settlements, carbonate rocks bearing admixture of the volcanomictic material occur in close association with volcanogenic–terrigenous beds. Carbonate sequence in the Cheremukhovo Settlement area is represented by thick limestone lentils intercalated with volcanic and volcanogenic–terrigenous members of the Sos'va Formation, and this section is close to that of the Vagran River area. In the area between the 2nd Severnyi Mine and Sos'va River, the Sarainaya Horizon includes an upper part of the Sos'va Formation representing a thick sequence of volcano-sedimentary rocks (Shatrov *et al.*, 1987). The maximum thickness of volcanics is recorded in the Mt. Kal'inskaya area. Southward, their thickness decreases, and in the area of the 3rd Severnyi Mine, these rocks wedge out. Simultaneously, the thickness of volcano-sedimentary rocks, mainly of tuffaceous conglomerates, increases in the same direction.

Different biomorphic and organogenic–detrital rock varieties of the Sarainaya Horizon are distinct among other rocks of the Petropavlovsk Formation owing to abundance and diversity of benthic fossil assemblages: crinoid, brachiopod–crinoid, ostracod–brachiopod–crinoid, coral, sponge, and algal.

Most typical of algal limestones and of their fragments in carbonate breccia are laminated *Rothpletzella devonica* (Masl.) and nodular spherical forms *Renaclis devonica* Johns.; also abundant are *Wetheredella silurica* Wood, *Gervanella ducii* Wether, *G. conferta* Champ., and *G. maslovi* Shuysk. The rock-forming *Fistulella variabilis* Shuysk. occurs less commonly, mainly in the upper portion of the Severoural'sk Horizon, where the share of algal limestone slightly increases.

Stromatoporoids are rare: columnar colonies of *Parallelostroma* sp. are registered in exposure 8005, and scarce colonies of *Simplexodictyon decorum* (Khalf.), *Parallelostroma scabrum* Bogoyavl., and *P. multiplexum* Bogoyavl. are present in the upper part of the section in the vicinity of Kal'ya and Cheremukhovo settlements.

Characteristic of the horizon are abundant and well-preserved remains of coral thickets mostly represented by small bioherm-type structures reaching several square meters in size. These are commonly accumulations or colonies of one or two–three branching species, one of which is dominant. Thickets are most typical of granular skeletal–detrital limestones, whereas isolated colonies occur in calcareous sandstones and algal limestones. Main rock-forming species in coral limestone are tabulates *Cladopora actuosa* Yanet and *Egosiella alba* (Yanet), or tetracorals *Pseudoamplexus fascicularis* Soshk. and *Neomphyma originata* Soshk. They occur throughout the sequence and are abundant in all studied sections. Slightly less common are *Favosites kozlowskii* (Sok.), *F. lucidus* Yanet, *Parastratopora fallacis* Yanet, *Thamnopora sarmentosa* Yanet, *Spongophyllum halysitoides* Ether., *Lyrielasma prima* Shur., and others.

Brachiopods usually occur as small accumulations or as individual specimens in organogenic–detrital limestones and calcareous sandstones, less commonly, in algal varieties of these rocks. Their assemblage includes more than 20 species, most common among which are *Gypidula optata* Barr., *Anastrophia magnifica* Kozl., *Hebetoechia vagranica* (Khod.), *Spirigerina supramarginalis* Khalf., and *Atrypinella delicata* I. et M. Breiv.

Ostracods *Aparchites koneprusiensis* Pr. et Snajdr., *Bashkirina elongata* Pol., *Bairdia michailensis* Zenk., *Praepilatina angulata* Zenk., and *Longiscula chorda* Zenk. are mostly confined to sandy rock varieties. They are also abundant in black, poorly cemented calcareous sands typical of the lower part of the horizon (exposure 8019, interval of 530–570 m). Diversity of ostracod assemblages is generally low (about 15 species); many of species are relatively rare.

Crinoids are substantial, frequently principal components of organogenic-detrital limestones and calcareous sandstones. Locally, their detritus is abundant in fine-grained calcareous or calcareous–silty matrix of breccia beds. Common are dispersed columnals and

small fragments of stems, but locally abundant cups, long stems, and root stalks of crinoids can be found. The crinoid assemblage consists of about 40 species. Pisocrinids *Parapisocrinus quinquelobus* (Bather), *P.* sp., *Pisocrinus* ex gr. *ubaghsi* Bouska, *P.(?) astericus* (Schew.), and *P.(?) costatus* (Schew.) are most widespread and associate with *Botryocrinus(?) mirandus* Yelt., *Eucalyptocrinites(?) ligatus* (Quenst.), *Salairocrinus textus* (J. Dubat.), *Pandocrinus pandus* Stuk., *Pentapetalocrinus* ex gr. *nieczlavensis* (Yelt.), *Peribolocrinus paludatus* J. Dubat., and *P. proximatus* J. Dubat.

Distribution of fossils in the Sarainaya Horizon shows that its upper part (about 100 m) yielding diverse organic remains differs from the remaining strata. The upper interval is marked by the first occurrence of the following fossil groups: crinoids *Parapernerocrinus bouskai* Milicina, *Salairocrinus textus* J. Dubat., *Pentagonocyclus filigerum* Schew., *Kaplunaecrinus rimosus* (Schew.), *Peribolocrinus proximus* J. Dubat., and *Cyclocetocrinus* ex gr., *rarigranulatus* Yelt. et Stuk.; corals *Thamnopora faceta* Yanet, *Th. solida* Dubat., and *Lyrielasma prima* Shur.; ostracods *Bairdiocypris* aff. *biesenbachi* (Krömm.), *Bairdiohealdites karcevae* (Pol.), and *Praepilatina praepilata saumica* Zenk. In the area between the 2nd Severnyi and Cheremukhovo settlements, the species mentioned above occur together with very specific thin accumulations of cylindrical sponges and single colonies of stromatoporoids *Simplexodictyon decorum* (Khalf.), *Parallelostroma scabrum* Bogoyavl., and *P. multiplexum* Bogoyavl., but stromatoporoids are extremely rare in the Sarainaya Horizon, e.g., they are missing from its stratotype.

The listed species confined to the upper ca. 100 m of the horizon contribute to paleontological characteristics of the entire horizon. The upper interval of its section is easily recognizable on the basis of fauna composition. It is well traced in the Vagran River basin (exposure 8020, Well 879^g) and in areas of the 3rd Severnyi Mine (exposures 8016, 8017), Kal'ya (Well 1205^g), Novaya Kal'ya (Wells 502^g, 1207^g, 1209^g), and Cheremukhovo (Wells 783^g, 1200^g).

Sauma Horizon

The stratotype of the Sauma Horizon is exposed along the eponymous river, in the outcrops located 800 m upstream of its mouth and along the left-hand bank of the Severnaya Toshemka River opposite the Sauma River mouth (Breivel' *et al.*, 1976; *Biostratigrafiya i fauna...*, 1977). In the vicinity of Severoural'sk, deposits of the horizon are referred to the Petropavlovsk Formation. Rocks directly underlying the bauxite deposits, i.e., limestones with *Karpinskia vagranensis* (Khodalevich, 1938) and Serov Horizon (Streis, 1951), have never been considered as members of the Petropavlovsk Formation, the reference section of which was originally distinguished in exposures along the left bank of the Vagran River. Later investigations by the paleontological–stratigraphic team revealed that fossils occurring there are also characteristic of underlying rocks. Now, the Sauma Horizon is defined in the range significantly exceeding that of the Serov Horizon, and its lower boundary is detected at the site located westward and downstream of the Vagran River.

The southern exposure of the Sauma limestones is located at the bank of the Vagran River, downstream of the Konovalovka River mouth. The continuous section of the horizon is exposed along the right-hand bank of the Vagran River opposite Severoural'sk (exposure 8020, interval of 590–1740 m). In the left-hand bank of the Vagran River, exposures of limestones characterize the upper part of the Sauma Horizon. Well 879g penetrated the entire Petropavlovsk Formation at the site located 400 m downstream of the old natural mouth of the Sarainaya River. The limestone thickness measured along the Vagran River is 450–500 m. Farther to the north, rocks of the Sauma Horizon are recovered in quarries and by deep wells. The thickness of the horizon near the 3rd Severnyi Mine is similar to that observed along the Vagran River. East of the Kal'ya Settlement (Well 7214), it decreases to 300 m and to 200 m in the Novaya Kal'ya area (Well 6783). In the Mt. Kal'inskaya area, the Sauma Horizon is missing (Wells 502^g, 1207^g, 1209^g).

The boundary between the Sarainaya and Sauma horizons is marked by transition from detrital patchy limestones with abundant benthic fossils to pale *Fistulella* limestone relatively impoverished in organic remains. This transition is particularly distinct along the right-hand bank of the Vagran River (exposure 8020, 590 m; Well 879^g). In the 3rd Severnyi Mine area, eastward of the Kal'ya Settlement, and in the eastern Novaya Kal'ya area, the bedded detrital limestones are gradually replaced by algal limestones. A sharp lithological transition between units was observed in the western Kal'ya area (Well 1205^g). The boundary between the Sarainaya and Sauma horizons marks faunal changes, i.e., appearance of new species despite a fairly high content of transient taxa.

The Sauma Horizon is composed of carbonate rocks, mainly of pale massive *Fistulella* and less common detrital–algal and algal–skeletal–detrital limestones. The characteristic feature of algal limestones from the Sauma Horizon is abundance of cloddy material, particularly in their lower beds (exposure 8020, interval of 800–850 m). This feature differs these rocks from limestones of the Sarainaya Horizon, where the main rock-forming organisms are represented by tubular, filamentous, and dendrite algal thalluses. Higher in the section, the Sauma Horizons composed of more granular and heterogeneous rocks, some of which represent synsedimentary breccias with fragments cemented by cloddy algal or micritic pink matrix displaying an insignificant admixture of muddy material.

Brecciated rocks are abundant in the Kal'ya area (Well 1205*g*, depth interval of 0–181 m).

In the Vagran River area, rocks directly underlying the ore horizon enclose lithoclasts of algal limestone, separate algal clods, and skeletal, mostly crinoid fragments, all cemented by micritic matrix. Encrustation structures are small.

Rock-forming organisms are represented by *Fistulella variabilis* Shuysky of doubtful taxonomic affinity, and also by algae *Renalcis devonicus* Johns. and *Ikella vermicularis* Shuysky. Other algae species are rare. Algal limestones bear rare benthic remains more abundant in the upper part of the horizon. Scarce stromatoporoids are represented only by massive and cylindrical colonies of *Densastroma moldavancevi* Yavor., *Parallelostroma multiplexum* Bogoyavl., and *P. scabrum* Bogoyavl. Corals occur more frequently. Among ten determined tabulate species, most abundant are *Favosites intricatus* Barr., *F. shirictensis* Chekh., and *Riphaeolites virgosus* Yanet. They associate with solitary tetracorals *Spongophylloides perfecta* (Wdkd.), *S. improcerus* Shur., and *Tryplasma aequabilis* Lonsd. Colonial *Pseudoamplexus quadripartitus* Soshk. and *Neomphyma simplex* Vag. are confined to upper limestone beds. Most common brachiopod species are *Gypidula nucalis* Khod., *G.* cf. *nux* Khod., *Ivdelinia pseudoivdelensis* (Khod.), *I. procerula* (Barr.), *Clorinda pseudolinguifera* Kozl., *Clorindina toschemkaensis* Khod., *Atrypinella losvensis* Khod., *A.* cf. *biloba* Khod., *Karpinskia vagranensis* Khod., *Septatrypa zelia* (Barr.), *Merista solita* I. et M. Breiv., and *Quadrithyris minuens* (Barr.). Some of these species occur also in the Sarainaya Horizon, whereas *Clorindina toschemkaensis, Atrypinella barba, Karpinskia vagranensis*, and *Quadrithyris minuens* are characteristic of the Sauma Horizon only. Ostracods are less abundant than in the Sarainaya Horizon, but their composition is almost similar in both units. Missing in the Sauma Horizon are genera *Silenis* and *Longiscula*, but characteristic species *Bairdiocypris* aff. *biesenbachi* Kromm., *Baschkirina indistincta* Zenk., and *B. retusa* Pol. appear at this level. Crinoids are also less abundant than in the Sarainaya Horizon. They occur locally in algal limestones and are concentrated only in the surface layer of the Krasnaya Shapochka bauxite deposit (exposure 333/62). Most characteristic among 15 determined species are *Cupressocrinites (Tetraptocrinus) ovatus* Schew., *Eucalyptocrinites(?) ligatus* (Quenst.) *Agatocrinus* cf. *aculeatus* (J. Dubat.), *Kuzbassocrinus bystrovi* Yelt., *Salairocrinus* ex gr. *jucundus* J. Dubat., and *Grossotocrinus* cf. *gradatus* (Yelt.).

The area between Mt. Kal'inskaya and the Sos'va River is occupied by a thick unit of carbonate–terrigenous rocks of the Sarainaya Formation overlying the Sos'va Formation. We refer this unit to the Sauma Horizon. It should be noted that the Sarainaya Formation is not coeval with the Sarainaya Horizon.

The Sarainaya Formation was first defined by S.M. Andronov (unpublished reports of 1942–1943), who considered it as a unit underlying the Petropavlovsk Formation. Later, the Sarainaya Formation was attributed to facies analogue of the lower part of the Petropavlovsk Formation (Peive, 1947; Streis, 1951; Andronov, 1961). Andronov selected the succession of dark bedded limestones that is exposed on both sides of the Sarainaya River in the 2nd Severnyi area of the Krasnaya Shapochka deposit as the type section of the Sarainaya Formation. He also referred to this formation the dark limestones exposed in the drainage trench of the former Boksity Settlement (now Severoural'sk), calcareous sandstones, small-pebble conglomerates with interbeds of brecciated gray and pink limestones exposed in the Kal'ya deposit, as well as dark limestones and sandstones studied in the vicinity of the Cheremukhovo Settlement. Our investigation revealed that the dark limestones of the last site and at the left-hand bank of the Vagran River (territory of Severoural'sk) correspond to the Sauma Horizon. In other areas, rocks referred by Andronov to the Sarainaya Formation are highly variable. For instance, in the area between the 2nd Severnyi Mine and Novaya Kal'ya Settlement, carbonate rocks are variegated, largely represented by breccias, and correspond, according to tetracoral and crinoid assemblages, to the Sarainaya Horizon. Similar rocks are also known in the Cheremukhovo area (Well 1200*g*, interval of 1162–1295 m; Well 783*g*, interval of 1541–1780 m) as constituents of the Sos'va Formation overlain by the Sarainaya Formation. In the entire area between the 2nd Severnyi Mine and Sos'va River, rocks of the Sarainaya Formation are transitional in composition between volcanogenic and volcano-sedimentary deposits of the Sos'va Formation and massive limestones of the Petropavlovsk Formation. The upper boundary of the Sos'va Formation corresponds here to the top of the Sarainaya Horizon.

In the Cheremukhovo area, rocks of the Sarainaya Formation are exposed along the Kedrovyi Creek and recovered by mine quarries, or penetrated by prospecting and hydrogeological wells.

In the section recovered near the Cheremukhovo Settlement (Wells 439^{g}, 441^{g}), the Sos'va Formation is overlain by the following successive beds:

(1) Limestones, dark–gray to gray, mainly skeletal–detrital-muddy, locally intercalated with interbeds of clayey limestone and calcareous sandstone. The latter is composed of algal particles and rare skeletal fragments submerged into the micritic matrix. These rocks alternate with members (to 7–8 m thick) composed of volcanogenic–terrigenous sandstones and siltstones; locally, matrix in these rocks is calcareous.

Some limestone beds enclose abundant corals and stromatoporoids, the components of biostromes and local coral thickets. Preponderant are massive and cylindrical stromatoporoid colonies of *Simplexodictyon decorum* (V. Khalf.) and *Parallelostroma muti-*

plexum Bogoyavl.; locally abundant are tabulates *Favosites socialis* Sok. et Tes. and *F. kozlowskii* (Sok.) associated with tetracorals *Spongophylloides perfecta* Wdkd. and with accumulations of brachiopods *Protathyris praecursor* Kozl. Clayey–calcareous rock varieties are rich in ostracods *Aparchites convexus* Zenk., *Pibylites (Parapibylites) sulcifer* Zenk., *Cavellina (Invisibilia) saraica* Zenk., *Healdianella asymmetrica* Zenk., *Bairdiocypris intermedius* Zenk., *Orthocypris tschumyschensis* Pol., *Bairdiohealdites karcevae* (Pol.), and *Baschkirina elongata* Pol.

The unit thickness is 70–95 m (Well 441, interval of 200–281 m; Well 439, interval of 210–320 m).

(2) Alternating beds of dark to dark–gray bedded limestone and pale massive limestone. Usually, the rocks are cloddy, less commonly, skeletal–detrital–algal with the recrystallized cloddy matrix. They commonly enclose the coral-stromatoporoid thickets; their bedded varieties bear abundant ostracods. The fauna is similar to that in the previous unit; it consists of the same stromatoporoid, coral, brachiopod, and ostracod species. In addition, single specimens of *Soshkinella columnaris* Zhelt. and *Lanceomyonia borealiformis* (Siem.) are detected.

The thickness is up to 110 m (Well 441^{g}, interval of 70–200 m; Well 439^{g}, interval of 70–210 m).

(3) Limestones, light–gray, less commonly, gray to pinkish, mainly massive and, sometimes, bedded. Under the microscope, they can be identified as cloddy algal limestones often bearing Cyanophyceae. The algal cloddy matrix is variably recrystallized. The skeletal detritus is rare; skeletal fragments experience two types of textural transformation: disintegration and micritization. Benthic fossils are scarce and mainly confined to bedded rock varieties. Faunal groups are identical to those in two underlying units, but their representatives occur as single specimens. Characteristic species are *Parallelostroma multiplexum* Bogoyavl., *Spongophylloides perfecta* (Wdkd.), and *Protathyris praecursor* Kozl.

The thickness is up to 110–120 m. Higher in the section, there is an ore horizon.

The total thickness of the succession is 300–350 m.

The considered succession is peculiar among carbonate rocks of the Lochkovian Stage, as its rocks mostly are cloddy in texture, and principal in biocoenosis are stromatoporoids, ostracods, and some species tabulates and tetracorals. The above section (Wells 439^{g} and 441^{g}) reveals transition from bedded carbonate varieties with interbeds of volcano-sedimentary rocks to pure massive limestones. This sedimentation trend is also detected farther to the east of the site, where the lower member thinner, whereas the second and, particularly, the third one are thicker. Cores from wells drilled east and south of the Cheremukhovo Settlement, and rock samples from quarry sections yielded fossils missing from assemblages from Wells 439*g* and 441^{g}.

On the territory of Severoural'sk, gray and dark–gray cloddy limestones from ditches (exposures 8170, 8171) yielded the following fossils: stromatoporoids *Parallelostroma multiplexum* Bogoyavl., *P. scabrum* Bogoyavl., *Simplexodictyion decorum* (V. Khalf.), and *Actinostroma fedorovi* Yavor.; tabulates *Favosites* aff. *lucidus* Yanet; tetracorals *Spongophylloides perfecta* (Wdkd.) and *Tryplasma aequabilis* Lonsd.; ostracods *Aparchites messleriformis* Pol., *Pibylites (Parapibylites) incertus* Zenk., *Cavellina (Invisibila) kamyshenkiensis* Pol., *Newsomites notabilis kusnetzkiensis* Pol., *Orthocypris tschumyschensis* Pol., and *Bashkirina elongata* Pol. By lithology and faunal assemblage, the described rock sequence corresponds to that in the Cheremukhovo area, most likely to its lower half.

We consider the entire rock sequence in the Cheremukhovo area (the Sarainaya Formation and uppermost part of the Petropavlovsk Formation between the Sos'va Formation and ore horizon) as a facies analogue of the Sauma Horizon. This standpoint was included as a special opinion in the explanatory note to stratigraphic schemes of the Urals (*Ob'yasnitel'naya zapiska...*, 1994), whereas in the schemes proper (*Stratigraficheskie skhemy...*, 1993), and in different publications (Dubatolov *et al.*, 1968; *Biostratigrafiya i fauna ...*, 1977; Bogoyavlenskaya, 1979), this sequence is correlated with the Sarainaya Horizon.

Our standpoint is supported by the following considerations:

(1) The Sarainaya formation rests upon the Sos'va Formation, whose upper limestone layers yielded fossils peculiar of the Sarainaya Horizon (Wells 783^{g}, 777^{g}, 1200^{g}): *Cladopora actuosa* Yanet, *Egosiella alba* (Yanet), *Neomphyma originata* Soshk., *Parapisocrinus ollula hlubocepensis* (Bouska), and *Podoliocrinus* cf. *nikiforovae* Yelt.

(2) The thickness of limestones in the Sarainaya Horizon stratotype section is 270–300 m. When limestones from the Cheremukhovo area are referred to the Sarainaya Horizon, the total thickness of carbonate rocks in this case will be up to 550 m and more (Well 783^{g}). In addition, the horizon also includes volcanogenic and volcano-sedimentary rocks of a great thickness (700–1000 m).

(3) The sequence has stratigraphic position similar to that of limestones of the Sauma Horizon penetrated by wells in the Kal'ya and Novaya Kal'ya areas located eastward along the horizon strike.

(4) As far as lithology is concerned, the algal cloddy limestones of the Cheremukhovo sequence are closer to algal limestones of the Sauma Horizon of the Vagran River area than to detrital and organogenic–detrital limestones of the Sarainaya Horizon. In the section of the Sauma Horizon in the Vagran River area (exposure 8020, interval of 800–850 m), most common are cloddy textures.

(5) The faunal assemblage from the Cheremukhovo sequence includes abundant endemic species confined

to dark limestone varieties. Remarkable among them are brachiopod (*Protathyris praecursor* Kozl.) and ostracod species. The ostracod assemblage is of peculiar composition. Preponderant among ostracods are representatives of the *Pibylites, Clavofabelina*, and *Cavelina* genera missing from the Petropavlovsk Formation in the Vagran River area and associated with *Bairdiocypris* and *Healdianella* forms differing from those occurring in reefal limestones. On the other hand, the Sarainaya Formation lacks representatives of the *Libumella, Praepilatina, Saumella*, and *Tricornina* genera typical of limestones from the Vagran River sections.

Fairly abundant in the faunal assemblage are transient species recorded in both the Sarainaya and Sauma horizons of the Vagran River area (stromatoporoids and some tabulates). Particular species occur separately in the Sarainaya Horizon (rare), or only in the Sauma Horizon. In the latter, they are represented by tetracorals *Soshkinella columnaris* Zhelt (Well 441[g] near the Cheremukhovo Settlement; exposure 8020, 610 m; Well 879[g] in the Vagran River area). Also important is the first occurrence of crinoid species with quadrilobulate central stem canal (*Cupressocrinites* sp.) in the Cheremukhovo sequence (exposure 4141, 783[g]), which are also detected in the Sauma Horizon of the Vagran River area (exposures 333/62, 336/62, and 8020, 850 m).

The aforesaid allows limestones underlying the Subrov Formation of bauxite-bearing rocks to be referred to the Sauma Horizon within the entire area between the Vagran and Sos'va rivers. An exception is the Mt. Kal'inskaya area, where the Sauma Horizon is missing and rocks of the Karpinsk Horizon rest directly on the Sarainaya Horizon (Sos'va Formation).

Rocks of the Severoural'sk, Sarainaya, and Sauma horizons distinguished in the vicinity of Severoural'sk are traceable also northward, in the Ivdel' area. In connection with the problem of relationships between the Sarainaya and Petropavlovsk formations, the sections exposed along the Severnaya Toshemka, Tosem'ya, and Vitim-Yatiya rivers are of prime significance. In these sections, tuffstones and tuffaceous conglomerates resting on limestones of the Sarainaya Horizon underlie a bedded member of gray to dark–gray limestones 150–170 m thick. The member yielded *Parallelostroma multiplexum* Bogoyavl., *Amnestostroma fedorovi* (Yavor.), *Thamnopora faceta* Yanet, *Spongophylloides perfecta* (Wdkd.), *Protathyris praecursor* Kozl., *Clavofabellina abunda minor* Pol., *Cavellina (Invisibilla) kamyshenkiensis* Pol., *Healdianella asymmetrica* Zenk., and others. The assemblage is similar to that from the Sarainaya Formation of the Cheremukhovo area. Higher in the section, gray and dark-gray limestones are replaced by pale massive algal limestones, locally bedded, displaying microcloddy texture, and containing similar, but less abundant fossils. These rocks are overlain by massive detrital limestones of the Vizhai Horizon of the Pragian Stage. Rocks of the last level bear the following fossils: corals *Riphaeolites* cf. *vijaicus* Yanet and *Lyrielasma petshorense* (Soshk.); brachiopods *Sphaerirhychia* cf. *vijaica* (Khod.); and abundant ostracods *Microcheilinella malobatschatskiensis uralensis* Zenk., *Bairdiocypris ivdelensis* Zenk., *Bekena borealis* Zenk., and others. The stratigraphic position of the Sarainaya Formation analogues in this sections is in favor of correlation between the Sauma Horizon and limestone sequence of the Sarainaya Formation distinguished in the Cheremukhovo area.

STAGES OF THE SUCCESSION FORMATION

Succession of horizons in the Severoural'sk section corresponds to particular sedimentation stages. Changes in sedimentation patterns resulted in gradual transformation of textural–structural rock properties and biocoenoses. Boundaries between stages mark particular events in the geological history, as well as changes in sedimentation settings and faunal communities.

The following main stages can be distinguished in the sequence of formations beginning from the Severoural'sk time.

Carbonate rocks of the uppermost Bobrovka Horizon indicate the onset of transgression and gradual transition from coastal to open shallow sea environments. Under these conditions, massive algal limestones of the Severoural'sk Horizon were formed. Abundance of algae probably decelerated development of benthic animals. The second half of the Severoural'sk time was marked by activation of hydrodynamics that is recorded in deposition of organogenic–detrital limestones among algal varieties and in the appearance of carbonate breccia. Areas located north of the Kolonga River accumulated the well-sorted polymictic sands, silts, and clays intercalated with rare gravel interbeds. In general, the Severoural'sk time was characterized by the relatively calm hydrodynamic conditions of sedimentation.

The transition to the Sarainaya time was marked by elevation of the erosion base and by accumulation of detrital, carbonate, and volcano-sedimentary rocks within the entire study area. Carbonates accumulated under conditions of the differentiated bottom topography and under influence of currents and wave erosion. Characteristic of the stage was formation of synsedimentary breccias, calcareous sandstones, and detrital limestones, often displaying the intense pigmentation by Fe-hydroxides. Corals and crinoids were significant rock-forming components. In the area between the 2nd Severnyi Mine and Sos'va River, the thick sequence of tuffaceous conglomerates, tuffstones and sandstones intercalated with basalt flows was formed at that time. Volcanics are most thick in the area near the Mt. Kal'inskaya, which probably represented a volcanic center. Carbonate accumulation was intermittent that is

particularly evident from the upper part of the sequence. The thickness and abundance of carbonate layers increases southward and northward of the Mt. Kal'inskaya.

The Sauma time corresponded to the onset of uniform conditions, when carbonate, mainly algal sedimentation was characteristic of the entire study region. Slightly peculiar environments existed only in the area between the Mt. Kal'inskaya and Sos'va River, where cloddy limestones were accumulated. They represent sediments of the most shallow type, as compared to other algal limestones, and accumulated in a basin zone influenced by active wave hydrodynamics. Sedimentation rates decreased. Differences in composition of benthic communities resulted not only from changes in sedimentation settings, but also from a partial basin isolation by a barrier located in the Mt. Kal'inskaya area.

All researchers assigned limestones of the Petropavlovsk Formation to reefal deposits, because they display some peculiar features. The performed studies revealed peculiarities in the structure and evolution of the Petropavlovsk reef. Main rock-forming organisms were represented by calcareous algae, problematic *Fistulella* forms, locally, by corals. Rocks are massive (except those in the lower part of the Sarainaya Horizon), sometimes, detrital and displaying encrustation structures. Limestones are light colored and lacking the terrigenous fraction (the content of insoluble residue is 1–2%). The thickness of the sequence is considerable. The reef included a complex facies succession, e.g., the fore-reef and back-reef facies. Nonetheless, deposits of the Petropavlovsk Formation do not correspond completely to the modern reef model, because the facies of lagoons and straits that separated the reef from the shore are missing. Facies of this kind are also unknown in other areas of the eastern slope of the Urals, where analogues of the Petropavlovsk Formation are distinguished.

CONCLUSION

The data obtained permit us to clarify positions of boundaries between the Severoural'sk, Sarainaya, and Sauma horizons in the continuous carbonate succession of the Petropavlovsk Formation, to elucidate their lithological and paleontological characteristics, and to trace these horizons northward and southward (Fig. 3).

To the north of the Kolonga River, the lower boundary of the Severoural'sk Horizon corresponds to the Sos'va Formation base. In the area between the Bol'shaya Kal'ya and Sos'va rivers, the lower boundary of the Sarainaya Horizon (correspondingly, the Silurian–Devonian boundary) is located inside the Sos'va Formation, approximately at the level of rudaceous tuffoconglomerate bearing fauna peculiar of the Sarainaya Horizon. The upper boundary of the Sarainaya Horizon corresponds to the top of the Sos'va Formation. Dark limestones and sandstones of the Cheremukhovo area (Sarainaya Formation) are correlative with the Sauma Horizon. Limestones underlying the Subrov Formation of bauxite-bearing beds are attributed to the Sauma Horizon in the entire area between the Vagran and Sos'va rivers, except for the Mt. Kal'inskaya area, where rocks of the Karpinsk Horizon rest upon deposits of the Sarainaya Horizon.

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New Data on Upper Cretaceous Deposits in the North of Moscow Region

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Abstract—New data on Upper Cretaceous deposits exposed to the north of Yaroslavl and to the southwest of Vladimir are considered in order to show that the Khot'kovo Opokas, dated back to the Turonian by some researchers, are actually younger and correspond to the Coniacian–Santonian level. The Turonian age is identified for underlying beds of carbonate clay, because they yielded cephalopods and radiolarians typical of this stage, and also an assemblage of early Turonian foraminifers. The beds are attributed to the Chernevo Formation. Holo- and parastratotype of this formation are described, and its stratigraphic range is outlined. New data suggest connections between the Turonian sea basins of the East European and West Siberian paleobiogeographic provinces.

Key words: Khot'kovo Opokas, Turonian Stage, Chernevo Formation, Moscow syneclise, calcareous clay, siltstone and sandy opoka beds, foraminifers, radiolarians.

The investigation history of Upper Cretaceous deposits in the Moscow syneclise began more than a century ago, but they are still inadequately studied. The Cenomanian age of the Varavino Group, whose sands overlie the erosion surface of relatively thick clay beds of the upper Albian Paramonovo Formation, is established here more or less confidently. However, the overlying terrigenous and siliceous deposits corresponding to the greater part of the Upper Cretaceous Series are doubtfully dated, and their stratigraphic range is so far under discussion (Kuzmenko *et al.*, 1997; Lavrova, 1971; Naidin, 1986).

The Upper Cretaceous deposits in question are exposed mainly to the north of Moscow within the sublatitudinal area of the Klin–Dmitrov Range and Yur'evskoe Opol'e, which extends from Klin and Solnechnogorsk in the west to Gavrilov Posad and Suzdal' in the east. Their small remnants after the Quaternary erosion are known near Kaluga, between Zaraisk and Mikhailov, in the vicinity of Chuchkovo and Shatsk (Ryazan' region), and in the Teplyi Stan district in the south of Moscow.

HISTORICAL INFORMATION

The first attempt to determine age of terrigenous–siliceous beds, which overlie Cenomanian deposits and have been termed later as the Khot'kovo Opokas (Nikitin, 1888), was that of Auerbach (1865), who studied the opoka and tripoli samples collected in the 15-m-deep excavation three miles away from Khot'kovo toward Sergiev Posad, where the Aleksandrovskaya (now Severnaya) railroad was under construction at that time. He detected fish remains (scale, teeth, and vertebrae) in the rocks, and also impressions of bivalves, those of inocerams included. Two species (*Inoceramus mytiloides* Goldf. and *I. brongniarti* Sow.) were identified among the latter, and this allowed Auerbach to correlate the Khot'kovo Opokas with marl beds of the Brongniarti-Plener in Germany and to date them back to the Cenomanian (Auerbach, 1865).

In the same year, Eichwald (1865) detected one more inoceram species *Inoceramus cuvieri* Sow. in these opokas and correspondingly attributed the rocks and underlying sand beds to the Turonian. In addition, he redefined *Inoceramus mytiloides* Mant. that was described by Auerbach as *Inoceramus lobatus* Münst (Eichwald, 1865–1868).

Twenty years later, Nikitin (1888) pointed out in his classical work *Traces of Cretaceous Period in Russia* that Trautschold (1872) detected the typical Cenomanian ammonite species *Schloenbachia varians* (Sow.) in sands of the Varavino Ravine. Studying his own collection of inocerams sampled in the Khot'kovo area,

this researcher defined among them *Inoceramus labiatus* (Schloth.), *I.* aff. *lobatus* Schlöt, and described new form *I. russiensis* Nik. As a result, he correlated the Khot'kovo Opokas with "the White Chalk of the Orlov and Kursk areas" considering both rock units as the early Turonian in age.

Later on, Pavlov (1900) correlated the Khot'kovo Opokas with the marl-and-chalk sequence of the Simbirsk and Saratov regions, and also with beds of opoka and tripoli exposed near Bryansk. Since faunas from these sediments turned out to be typical of the *Inoceramus brongniarti* Zone (Brongniarti-Plener) of northern Germany and included, in addition to the index species, the Coniacian forms *Inoceramus russiensis* Nik. and *I. involutus* Sow. associated with the late Turonian *Inoceramus cuvieri* Sow., Pavlov concluded that the Khot'kovo Opokas are not older than the late Turonian. He believed as well that the Emscherian transgression of the Coniacian–early Santonian time is recorded in middle courses of the Volga River but not in the Moscow and Vladimir regions, where the Khot'kovo Opokas must be thus limited in stratigraphic range by the Turonian. Afterward, Pavlov (1903) changed his mind and considered the range of the beds as corresponding to the uppermost Turonian and basal Emscherian (Coniacian), because "the Late Cretaceous seawaters penetrated far to the north at the Turonian–Senonian boundary time, when they flooded areas around Moscow" (p. 145).

Dobrov (1922) also argued for the Emscherian (Coniacian–lower Santonian) level of glauconite–clayey sand, sandstone, clay, and opoka beds of the Khot'kovo sequence using his own determinations of *Inoceramus lamarcki* Park., *I. russiensis* Nik., and *Actinocamax verus* Müll. In addition, he reported that the lower sandstone bed of the sequence is pitted with holes that resulted from cavernous weathering of pebbles and thus mark the basal erosion level in the Orlov and Kursk type sections (p. 20).

Later on, Dobrov (1929) retuned back to the viewpoint of Pavlov (1903) and suggested the Emscherian age only for the upper part of the Khot'kovo Horizon beginning from the base of the cavernous sandstone bed. He also attributed tentatively the lower portion of the horizon (3–4.5 m thick), where "fauna was inadequately studied," to the Turonian

Three years afterwards, Dobrov (1932), considering as problematic the age of basal sandstone beds of the Khot'kovo Horizon, included them, nevertheless, into the Cenomanian interval in his illustrative geological column. The overlying beds with glauconite and clay admixture (only 1.0–2.5 m thick in total) were attributed to the Turonian, and cavernous sandstones with impressions of *Inoceramus russiensis* Nik., *I.* cf. *involutus* Sow., *I.* ex gr. *cordiformis* Sow., *I. kleini* Müll., *I. lobatus* Schlöt., *I. lamarcki* Park., and *I. percostatus* were correlated with the Emscherian.

Dan'shin (1947) was of the same opinion and considered only the lower part of the Khot'kovo Opokas (below the cavernous sandstone bed distinguished by Dobrov) as corresponding to the Turonian. However, he increased thickness of this stage in the north of Moscow region from 4.5 to 10 m using data of Dobrov, who recovered *Inoceramus lamarcki* Park. out of the bed of glauconite–quartz sandstone in the Ten'tikovo Quarry.

One year later, Dobrov (1948) decided to be safer in his age estimates of the Khot'kovo Horizon as he wrote that "the Turonian Stage in the Moscow–Vladimir region is insufficiently studied in the paleontological aspect" (p. 297). In his opinion, the stage could span only the lower and minor part of the horizon, and "we may just state that it is not older than the Turonian" (p. 298). According to Dobrov, *Inoceramus labiatus* Schloth. described by Nikitin (1888) does not correspond to this species originally described by Schlotheim. As for *Inoceramus lamarcki* Park. mentioned by Dan'shin as one of typical middle Turonian forms, it was included by Dobrov in the list of Coniacian inoceramus, and this idea finds support in the recent work of Walaszyk (1992), who established that species comparable with *Inoceramus lamarcki* Park. were recurrent in the lower and middle Coniacian. We obtained the same result collecting inocerams from all marl beds of the Chernetovo Formation that is characterized by foraminifers of the *Gavelinella kelleri* Zone at the Lyubozhichi site, 46 km northeastward of Trubchevsk, the right-hand side of the Desna River. Walaszyk identified the collected forms as *Inoceramus* ex gr. *lamarcki* Park.

It should be noted as well that Dobrov considered the hypothetical Turonian strata as increasing in thickness toward the east. In his opinion, beds of opoka-like sandstone, clayey–micaceous sand, and sandy opoka that are typical of the Dmitrov area grade into beds of black sandy clay exposed in the Sergiev Posad, Yur'ev-Pol'skii, and Gavrilov Posad areas, and also into calcareous clay beds of the Kol'chugino area. He thought that these facies changes resulted from an insignificant but regular depth increase within the Ryazan–Kostroma depression of the past sedimentary basin.

The tendency to suggest the Turonian age for lower beds of the Khot'kovo Opokas resided in some publications of recent decades. For instance, Naidin (1986) attributed to the Turonian "the lower beds of the so-called Khot'kovo Opokas resting on Cenomanian sands and consisting of tripoli, clay, sand, and opoka-like sandstone" (p. 107). The Coniacian Stage corresponds, in his opinion, only to the upper part of the Khot'kovo Horizon beginning from "the peculiar, cavernous and silicified sandstone," i.e., to beds of clayey tripoli, sand, and sandstone yielding an assemblage of *Inoceramus russiensis* Nik., *I. percostatus* Müll., *I. involutus* Sow., and *I. kleini* Müll.

The last viewpoint contradicts however to results obtained during the medium-scale geological survey in

1960–1966 and summarized by Lavrova (1971) in the fourth volume of *Geologiya SSSR* (Geology of the USSR). In this publication, the Khot'kovo Opokas are dated back to the Santonian on the basis of radiolarian fauna studied by R.F. Smirnova, whereas the Turonian Stage is considered as corresponding to beds of calcareous–siliceous clay, sandstone, and calcareous sand known solely in the eastern part of the Klin–Dmitrov range. Fauna from these beds includes belemnites *Goniocamax inermedius* (Arkh.) and foraminifers. According to Smirnova, who studied microfauna from many borehole sections and summarized other published and unpublished data, the foraminiferal assemblage consists of the following species: *Spiroplectammina praelonga* (Reuss), *Gaudryina angustata* Akim., *G. serrata* Franke, *Arenobulimina minima* Vass., *Valvulineria lenticula* (Reuss), *Gyroidinoides nitidus* (Reuss), *Globorotalites multiseptus* Brotz., *Eponides belorussiensis* Akim., *Gavelinella* ex gr. *ammonoides* (Reuss), *G. moniliformis* (Reuss), *G. vesca* (N. Byk.), *Lingulogavelinella globosa* (Brotz.), *Brotzenella belorussica* (Akim.), *B. berthelini* (Kell.), *Cibicides polyrraphes* (Reuss), *Cibicidoides praeeriksdalensis* (Vass.), *Hedbergella holzli* (Hagn.), *Praeglobotruncana umbricata* (Morn.), *Globotruncana globigerinoides* (Brotz.), *Reussella turonica* Akim., *Sitella gracilis* (Vass.), *Eouvigerina regularis* (Kell.), *Tappannina eouvigeriniformis* (Kell.), and *Heterohelix globulosa* (Her.). In opinion of Smirnova, the assemblage is indicative of Turonian–Coniacian age of host deposits, but Lavrova, who reasoned that the assemblage lacks typical Coniacian forms *Stensioeina emscherica* Bar., limited its time range by the Turonian Age.

In addition to foraminifers listed by Smirnova, the beds in question bear radiolarians defined as *Cromyodruppa concentrica* Lipm., *Spongodiscus volgensis* Lipm., *Orbiculiforma impressa* (Lipm.), *Stylodictya delicatula* Lipm., and *Histiastrum* aff. *tetracantum* Lipm. (Lavrova, 1971). All these forms, except for *Stylodictya delicatula* Lipm. characteristic of the Turonian deposits only, are of a wide (Turonian–Maastrichtian) stratigraphic range.

TURONIAN DEPOSITS OF THE MOSCOW SYNECLISE

Generalizing available data on Upper Cretaceous deposits in the Moscow syneclise, we established that the calcareous clay beds underlying siliceous sediments of the Khot'kovo Horizon occur everywhere (except for the exaration zones) in the Kol'chugino and Yur'ev-Pol'skii areas of the Vladimir region. Moreover, small erosion remnants composed of this clay are known to the west of Suzdal', where the beds in question are exposed in a ravine near the village of Andreevskoe. Further to the north, calcareous clay beds were recovered in borehole sections to the south of Gavrilov Posad (Ivanovo region). In the Yaroslavl region, they are known near the village of Koshelevo, southwestward of Pereyaslavl'-Zalesskii. Borehole sections recovered at the Chernevo and Potanino sites (southward of Berendeevo and Pereyaslavl'-Zalesskii, respectively) appear to be thickest and most complete. Thin and fragmentary Turonian beds are known in the extreme west of the Vladimir region (Borehole 130 near the village of Lobkovo of the Aleksandrov district). Distribution areas of these deposits are schematically shown in Fig. 1.

Geological profiles (Figs. 2 and 3) illustrate relations of these clay beds with underlying and overlying strata. In accordance with the stratigraphic scheme that was authorized by the regional Interdepartmental Stratigraphic Commission on Central and Southern Russian platform (3.12.1992), the profiles show subdivisions of Cretaceous deposits characteristic of the Moscow syneclise, namely the Paramonovo Formation of the upper Albian, the Yakhroma and Lyamino formations of the Cenomanian, and the Chernevo Formation of the lower Turonian, which is overlain in places by the Potanino Beds of the lower Coniacian. Upward in the sequence, there is the Khot'kovo Horizon (opokas) consisting of the Zagorsk, Dmitrov, and Ten'tikovo formations crowned by the Godunovo Beds. These units are tentatively attributed to the undivided Coniacian–Santonian interval (Olfer'ev, 1986; Kuzmenko *et al.*, 1997). The Zagorsk Formation includes the Khot'kovo beds occurring under the cavernous sandstone and thus corresponding, in terms of Dobrov and Naidin, to the "Turonian." Next formation comprises the cavernous sandstone bed and overlying sands and aleurites. The third one consists of tripoli and opoka beds. In the most complete sections, the Ten'tikovo Beds of tripoli are overlain by the Godunovo Beds of sandstones. Among all stratigraphic units listed above and originally distinguished by Olfer'ev, only the Paramonovo, Yakhroma, and Lyamino formation have been characterized in full accordance with requirements of the Stratigraphic Code (Olfer'ev, 1986). Other units usable in practice as conditionally valid (Bragina, 1994; Kuzmenko *et al.*, 1997) have not been properly described, and we partially fill up the gap in this work.

We studied two reference sections, which can be considered as holo- and parastratotype of the Chernevo Formation. The chosen holostratotype corresponds to the depth interval of 37.3–47.0 m in the section recovered by Borehole 102 at the Chernevo site. This borehole drilled in 1996 by the Geocenter of Moscow duplicates Borehole 20 that recovered the thickest sequence of Turonian deposits and was drilled in 1964 by the Moscow Hydrogeological Expedition of GUTSR in the course of the medium-scale geological survey. We also investigated core samples from the neighboring Borehole 101.

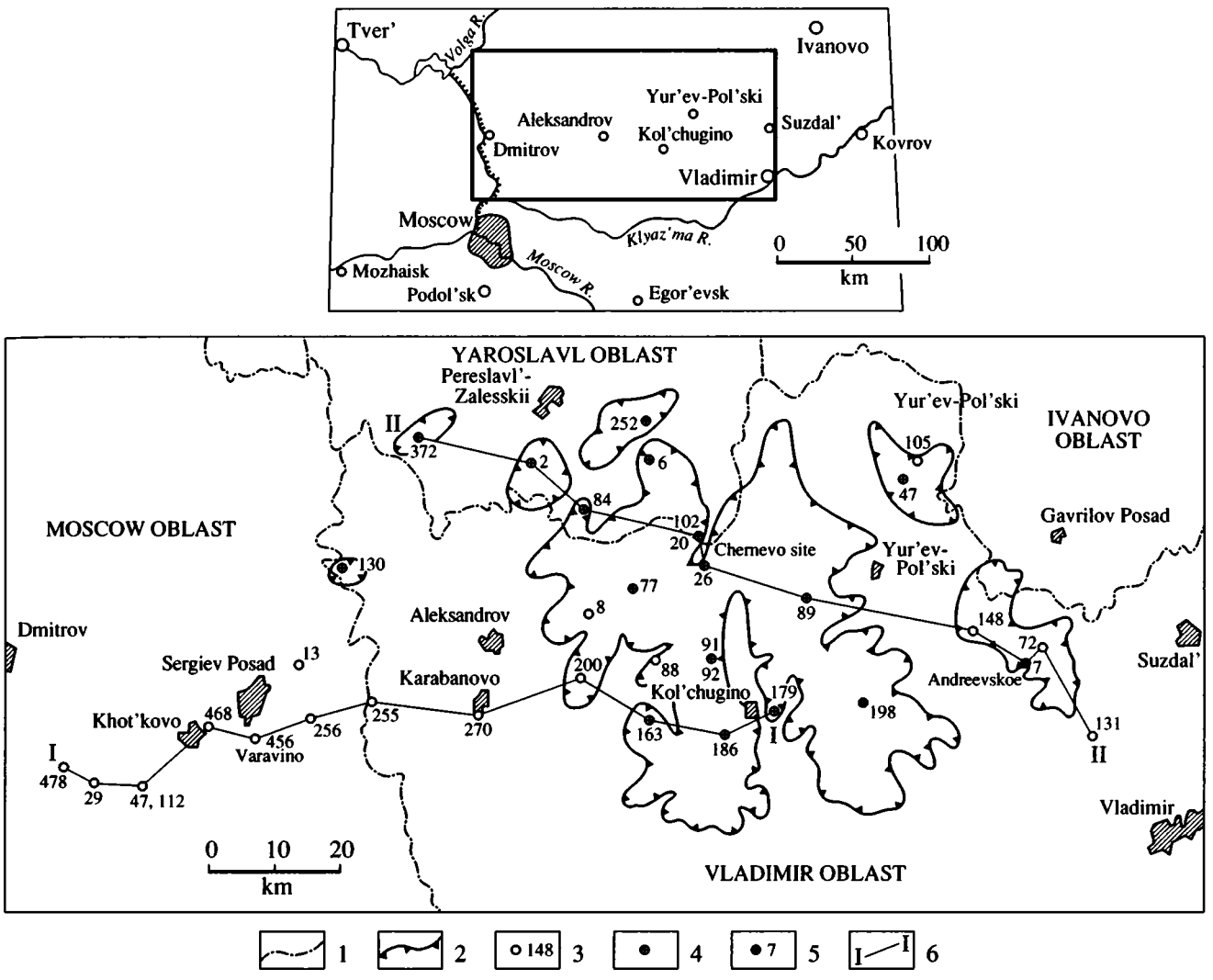


Fig. 1. Distribution areas of the Chernevo Formation: (1) administrative boundaries; (2) distribution limits of the Chernevo Formation; (3) borehole site and its number; (4) site that yielded foraminifers; (5) natural exposure and its number; (6) line of geological profile.

Borehole Section 102 (Chernevo Site, Yaroslavl region)

Boreholes 101 and 102 are located 0.5 km westward of the Chernevo Tract near the turn of road to the village of Bektyshevo and 13 km away to the east–southeast from Berendeevo of the Pereyaslavl'-Zalesskii area (Yaroslavl region). The boreholes recovered the sequence of Cretaceous deposits described below.

Under the base of morainal loam (depth level 20.6 m), there is recovered the following succession of stratigraphic subdivisions:

Dmitrov Formation

Bed 1 (20.6–25.3 m): dark gray, clayey, fine to coarse silt displaying signs of intense bioturbation. Narrow fucoids are filled with the darker clay mass enclosing thin laminae of light gray silt and gray silty clay. The clay content decreases downward, where the bed is light gray, but it grows again at the base. The bed shows gradual transition to underlying sediments.

Bed 2 (25.3–29.0 m): dark gray, clayey–micaceous fine silt almost lacking signs of bioturbation and showing a vague subhorizontal lamination. The clay content in individual laminae is variable, and some of them are lacking this material. The total content of pelitic material increases downward, and the fine silt grades first (below 27.0 m) into silty, and then (below 28.0 m) into pure clay. The lower boundary is very distinct.

Eight rock samples collected from both beds yielded abundant and well-preserved radiolarians. In terms of L.I. Kazintsova, who studied them, these fossils represent the *Orbiculiforma quadrata–Rhopalastrum attenuatum* assemblage (see Fig. 4 for the complete species

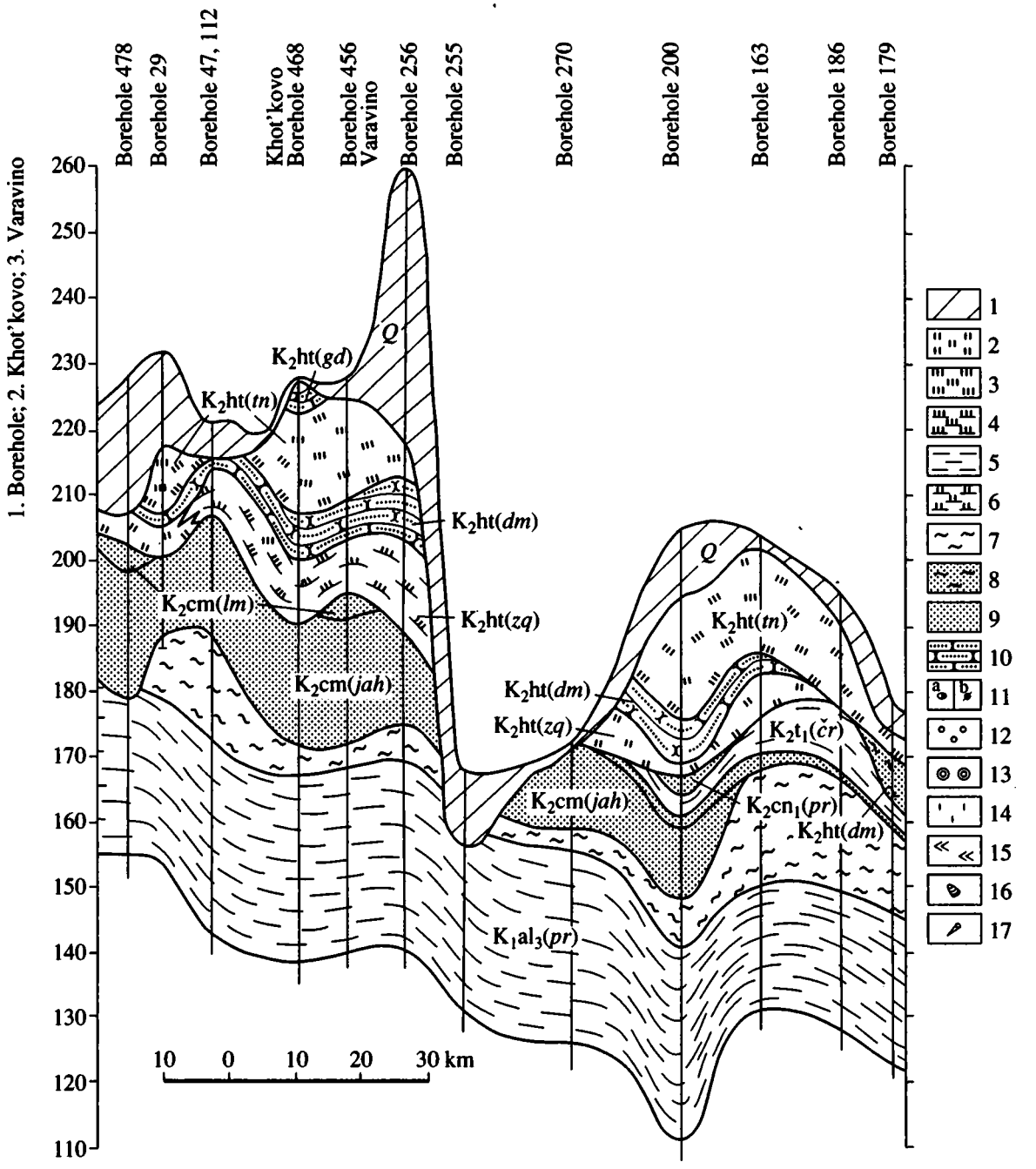


Fig. 2. Geological profile along the line I–I: (1) loam; (2) opoka; (3) tripoli; (4) clayey tripoli; (5) clay; (6) silicified clay; (7) silt; (8) sandy silt; (9) sand; (10) sandstone; (11) nodules (a), pebbles (b) of phosphorite; (12) cherty pebbles; (13) oolites; (14) carbonate admixture; (15) accessory mica; (16) inocerams; (17) belemnites. Indices of stratigraphic subdivisions: $K_1al_3(pr)$, upper Albian Paramonovo Formation; $K_2cm(jah)$, Cenomanian Yakhroma Formation; $K_2cm(lm)$, Cenomanian Lyamino Formation; $K_2t_1(cr)$, lower Turonian Chernevo Formation; $K_2cn(pt)$, lower Coniacian Potanino Beds; $K_2ht(zg)$, $K_2ht(dm)$, $K_2ht(tn)$, and $K_2ht(gd)$ indicate Zagorsk, Dmitrov, and Ten'tikovo formations, and also Godunovo Beds of the Khot'kovo Horizon (Olfer'ev, 1986; Kuz'menko *et al.*, 1997).

list). The assemblage includes many species in common with radiolarians characterizing the following units: the Santonian–Campanian *Dictyomitra striata* Zone of Slavgorod Formation in West Siberia (Lipman, 1962); the late Santonian *Prunobracchium crassum* and early Campanian *Patulibrachium petroleumensis* beds in the Usa River basin near the subpolar Urals (Amon, 1994); and the upper Santonian–middle Campanian strata in the Moscow region (Khot'kovo Group in terms of Bragina, 1994) and California (Pessagno, 1976).

Zagorsk Formation

Bed 3 (29.0–31.3 m, core recovery 2.0 m): gray hard opoka with an admixture of fine silt and mica flakes.

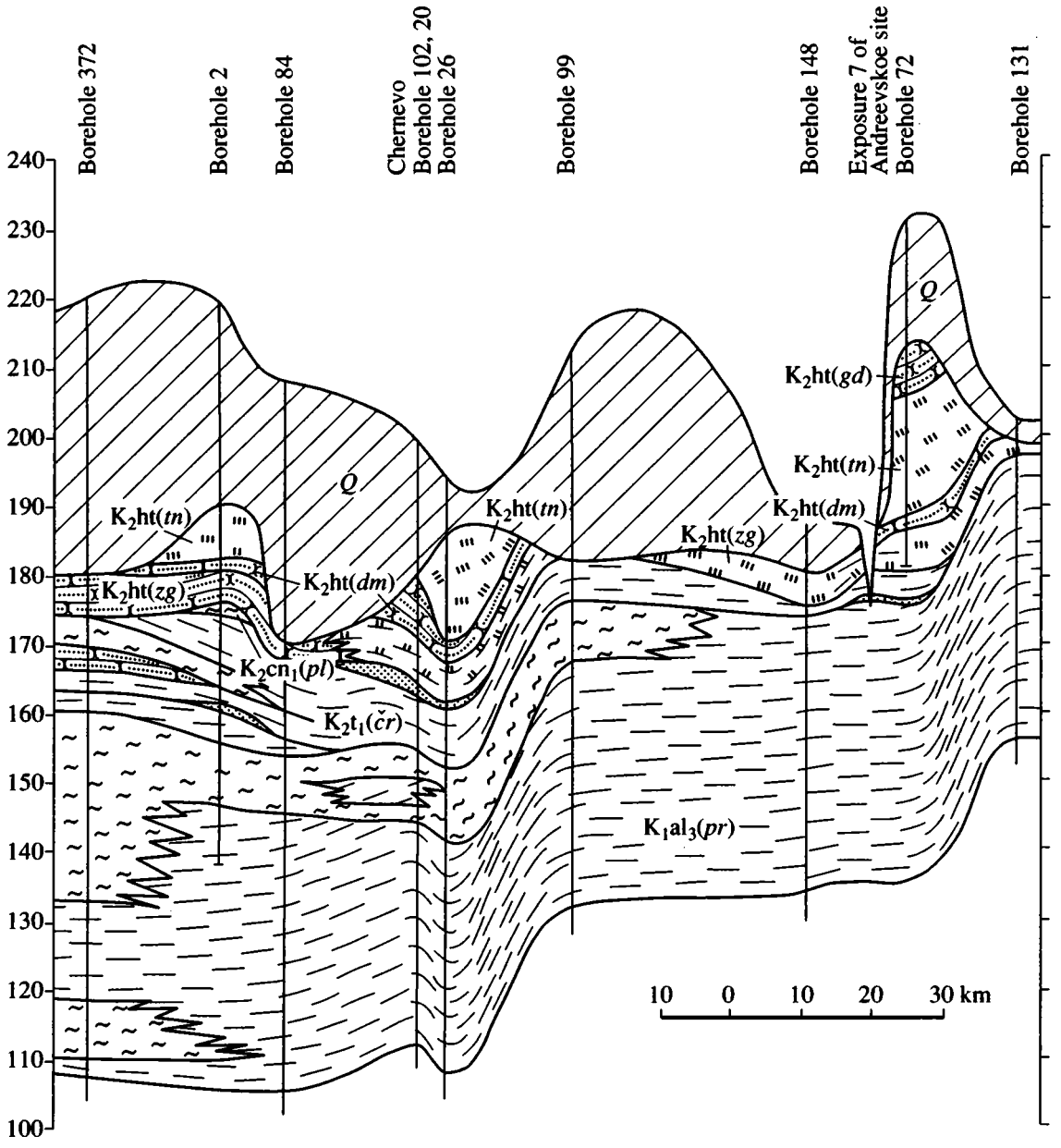


Fig. 3. Geological profile along the line II–II (symbols as in Fig. 2).

The well-outlined fucoids in the rock are filled with pale siliceous matter (opal) incorporating fine glauconite grains. The fine sandy fraction of glauconite appears closer to the bed base, where opoka is transformed into the dark and spotted sandstone with siliceous matrix. The lower boundary is distinct.

Bed 4 (31.3–34.0 m, core recovery 1.4 m): dark gray to black meager clay with admixture of very fine silt and mica flakes. The rock is massive, irregularly silicified, lacking lamination, and incrementally grading into the underlying bed.

Bed 5 (34.0–35.5 m, core recovery 0.9 m): gray to dark gray opoka with the irregularly distributed admixture of clay, silt, and mica flakes. The rock is locally hard, showing caverns after leached belemnite rostrums attributed by Naidin to the genus *Belemnitella* and also impressions of prismatic layer of inoceram shells. The transition between this and underlying bed is gradual.

Bed 6 (35.5–35.9 m, core recovery 0.4 m): dark gray meager, somewhat silty and micaceous clay displaying signs of intense bioturbation. Narrow and branching fucoids are filled with light gray silt free of clay admixture. The lower boundary is distinct.

Three samples from Bed 4 and one from Bed 6 yielded abundant radiolarians of the *Orbiculiforma vacaensis–Archaeospongoprunum rumseyensis* assem-

Geological section of Borehole 102, Chernevo site, Pereslavl'-Zaleskii area of the Yaroslavl region. Abs. mark of the borehole head is 200 m.

Fig. 4. Distribution of radiolarians in Cretaceous sediments recovered by Borehole 102 (symbols as in Fig. 2).

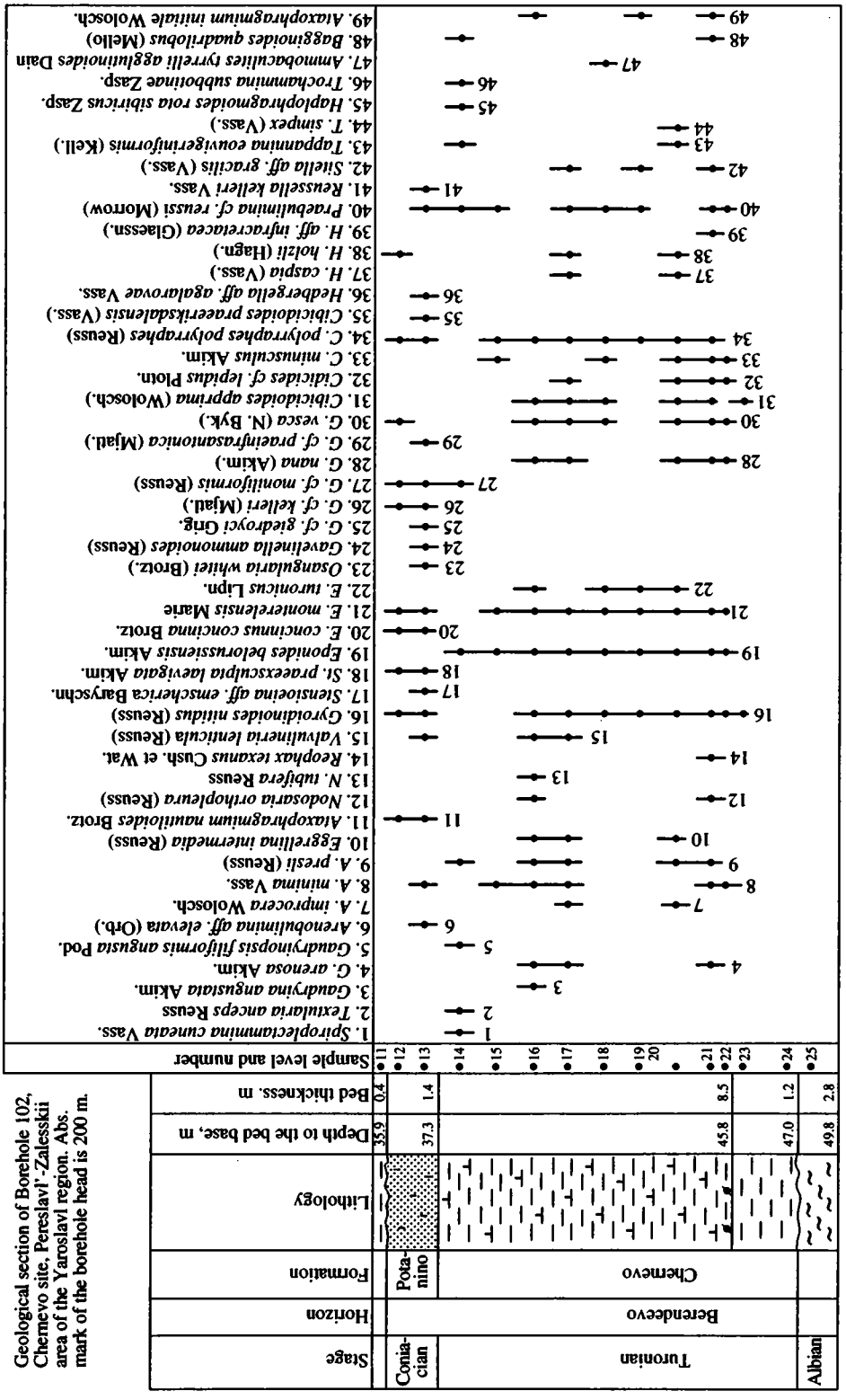


Fig. 5. Distribution of foraminifers in the Chernevo Formation and Potanino Beds, section of Borehole 102 (symbols as in Fig. 2).

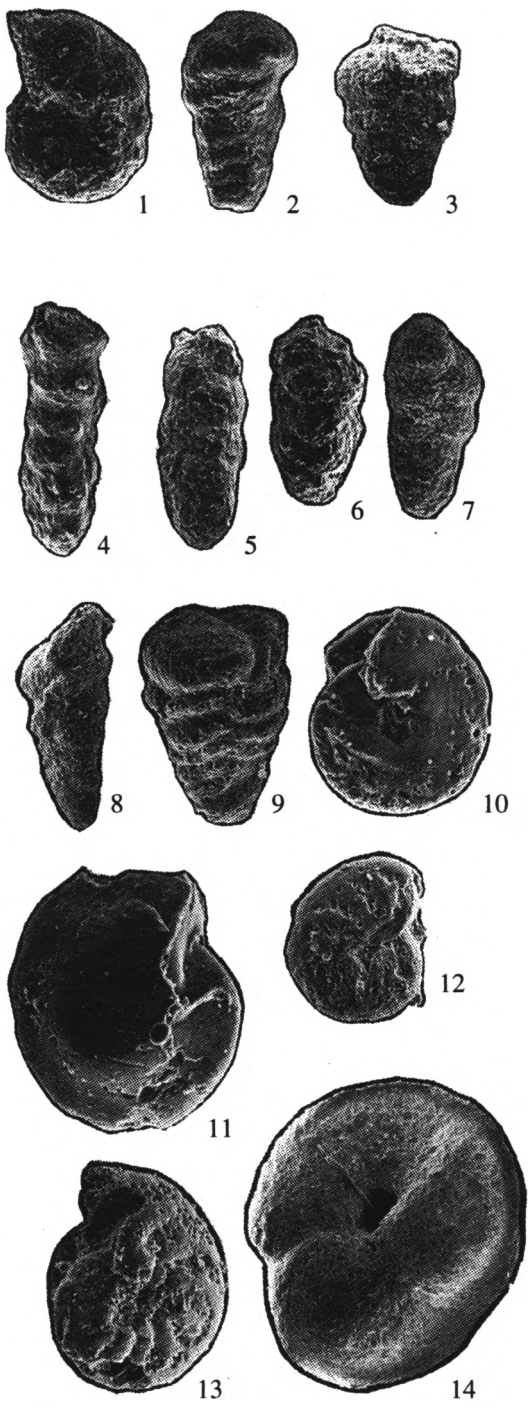


Plate I. Foraminifers from the Chernevo Formation (lower Turonian) and Potanino Beds (lower Coniacian), boreholes 101 and 102, Chernevo site of the Yaroslavl region: (1) *Haplophragmoides rota sibiricus* Zasp., ×80, Borehole 101, depth 38.0 m, lower Turonian; (2, 3) *Textularia anceps* Reuss, ×100, Borehole 101, depth 38.0 m, lower Turonian; (4–6) *Gaudryinopsis filiformis* (Berth.), ×80, Borehole 101, depth 38.0 m, lower Turonian; (7–9) *Verneuilinoides* sp., ×100, Borehole 101, depth 38.0 m, lower Turonian; (10, 11) *Cibicidoides praeeriksdalensis* (Vass.), ×200, Borehole 102, depth 37.0 m, lower Coniacian; (12, 13) *Gavelinella praeinfrasantonica* (Mjatl.), ×100, Borehole 102, depth 37.0 m, lower Coniacian; (14) *Ataxophragmium nautiloides* Brotzen, ×50, Borehole 102, depth 37.0 m, lower Coniacian.

blage. Some species of the assemblage are characteristic of the Coniacian radiolarian zones of *Ommatodiscus mobilis* in the Usa River basin near the subpolar Urals (Amon, 1994), and also of *Alievium praegallowayi* in California (Pessagno, 1976). At the same time, some other taxa are known from the early Santonian *Theocampe animula* assemblage of the Usa River basin (Amon, 1994) and from radiolarian faunas of the Dmitrov and Ten'tikovo formations studied by Bragina (1994) at the Khot'kovo, Varavino, and Psarevo sites of the Sergiev Posad area. According to Naidin (1964), *Belemnitella* forms usually appear beginning since the initial Santonian time, but we collected them also from the Coniacian deposits of the Voronezh anteclise. Thus, the Zagorsk Formation is likely corresponding to the major part of the Coniacian Stage coupled with the lower Santonian Substage.

Potanino Sands

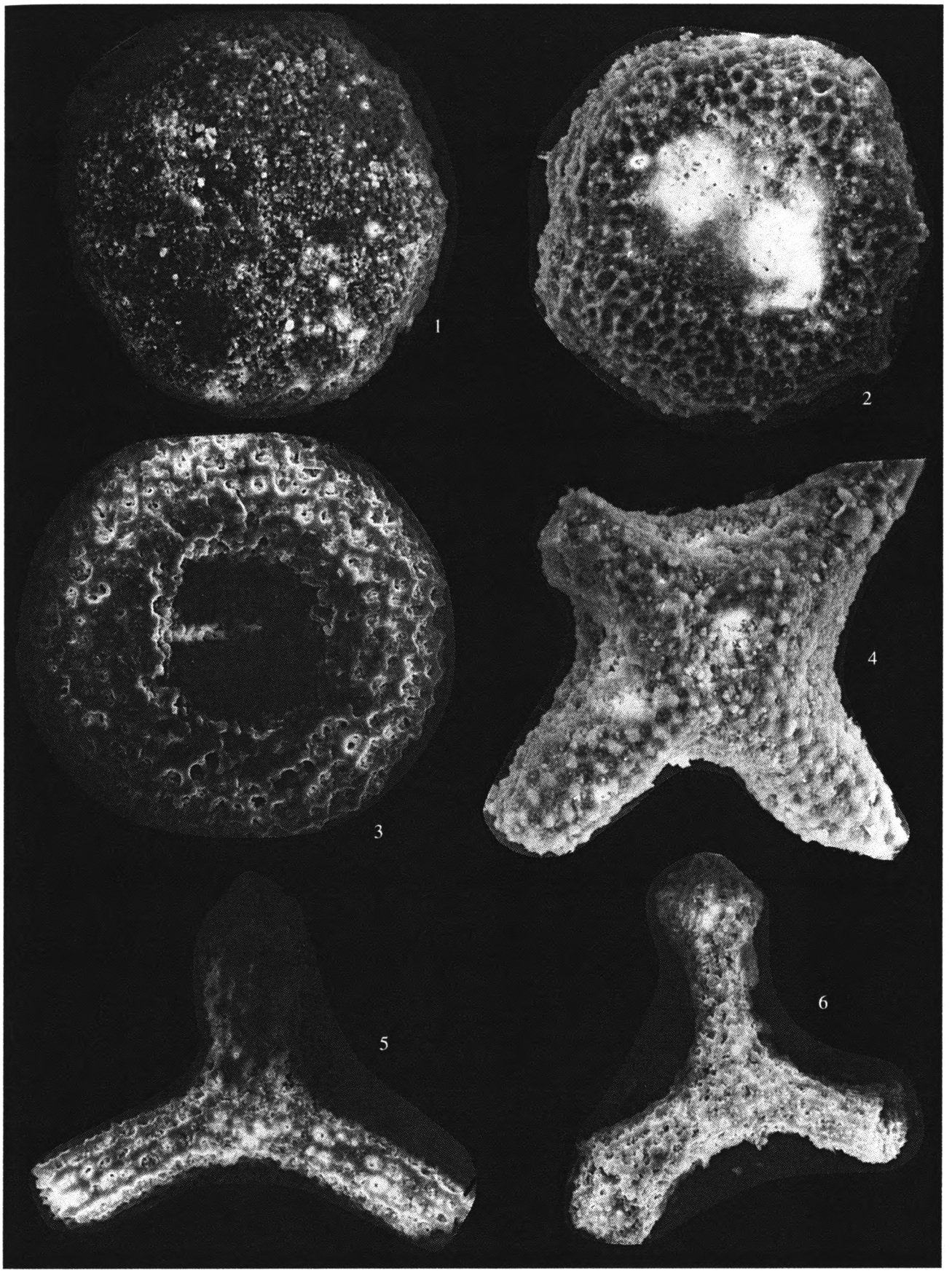
Bed 7 (35.9–37.3 m, core recovery 1.4 m): gray, fine-grained, calcareous quartz sand locally intercalated with dark gray interlayers of calcareous bioturbated clay. Tracks of large fossorial organisms (presumably of thalassinoids) are filled with sandy material of the bed. At the base, sand is cemented by calcareous matrix. The lower boundary is fairly distinct.

The bed yielded foraminifers characteristic of the lower Coniacian Substage (Fig. 5). The studied assemblage includes *Ataxophragmium nautiloides* Brotzen, *Osangularia whitei* (Brotzen), *Stensioeina praeexsculpta laevigata* Akim., *S.* aff. *emscherica* Bar., *Eponides concinnus* var. *concinna* Brotzen, *Gavelinella giedroyci* Grig., *G.* cf. *kelleri* (Mjatl.), *G.* cf. *praeinfrasantonica* (Mjatl.), *Cibicidoides praeeriksdalensis* Vass, and *Reussella kelleri* Vass. (Plate 1). These forms suggest the late Coniacian age for the upper beds of the Chernevo Formation. The listed foraminifers occur together with radiolarian species *Orbiculiforma vacaensis* Pess., whose distribution range corresponds to the Turonian–lower Santonian interval.

Chernevo Formation

Bed 8 (37.3–45.8 m, core recovery 8.5 m): meager silty–calcareous clay gray to dark gray in the wet state and ashy-gray after drying. The rock shows vague subhorizontal bedding owing to intercalated thin laminae of light gray glauconite-bearing silt. In local areas of intense bioturbation, the fucoids are filled either with light gray micaceous silt, or with whitish carbonate matter. A belemnite rostrum was recovered from the bed, and sandy phosphorite segregations were recognized at the base of the bed, which is somewhat blur.

The foraminiferal assemblage from the lower part of the bed (40.0–45.8 m) mostly consists of small secretory forms. Among the latter, there are *Eponides belorussiensis* Akim., *E. monterelensis* Marie, *E.* cf. *turonicus* Lipn., *Gavelinella nana* (Akim.), *G. vesca*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(N. Byk.), *Cibicides polyrraphes polyrraphes* (Reuss), *C.* cf. *lepidus* Plotn., *Praebulimina reussi* (Morrow), and *Tappannina eouvigeriniformis* (Keller). These and associated agglutinating species *Gaudryina arenosa* Akim., *Arenobulimina minima* Vass., and *A. presli* (Reuss) point to the early Turonian age of the host rocks (Plate 1). The upper portion of the bed is richer in shells of *Gaudryina arenosa* Akim. and corresponds to the first occurrence level of agglutinating foraminifers characterizing the *Gaudryinopsis filiformis angusta* Zone of West Siberia. The latter become dominant in the assemblage. It is also remarkable that, in addition to typical Turonian species occurring everywhere from the Mangyshlak and Caspian Sea coast to the Volhynia and Podolia, the assemblage includes forms characteristic of foraminiferal assemblages of the Kuznetsov Formation in West Siberia: *Gaudryinopsis filiformis angusta* Pod., *Textularia anceps* Reuss, *Trochammina subbotinae* Zasp., *Haplophragmoides rota sibiricus* Zasp., *Bagginoides quadrilobus* (Mello), and *Ammobaculites tyrrelli agglutinoides* Dain., which were unknown before in the Russian platform. In West Siberia, the *Gaudryinopsis filiformis angusta* Zone spans the lower and, partially, upper substages of the Turonian.

The radiolarian assemblage from the clay bed under consideration consists of 18 species (Fig. 5). Remarkable among them are typical Turonian forms *Cavidiscus fiskensis* Pess., *Cenosphaera magna* (Grig.), and *Stichocapsa pyramongata* (Grig.), as well as *Orbiculiforma multangula* Pess. (Plates II and III) that takes its origin in the Albian and terminates evolution in the Turonian. Some species, like *Dictyomitra napaensis* (Pess.), *Stylodictya placentalis* Lipm., *Conosphaera* cf. *mammilata* Lipm., *Spongodiscus maximus* Lipm., and *Ommatodiscus mobilis* Kozl., are known only from the Chernevo Formation of the Moscow syneclise. Other representatives of the assemblage (*Orbiculiforma cachensis* Pess., *O. vacaensis* Pess., *Spongodiscus volgensis* Lipm., *Cenosphaera minor* Lipm., *Dictyomitra multicostata* Zittel, and some others) proceed higher into the overlying Khot'kovo Horizon. The described assemblage is identical to the Turonian *Cenosphaera magna–Stichocapsa pyramongata* assemblage from the Kuznetsov Formation of the Cis-Urals region (Amon, 1985; Amon and Papulov, 1989). On the basis of foraminifers of the *Gaudryinopsis filiformis angusta* and *Pseudoclavulina hastata hastata* zones, the formation range was interpreted as corresponding to the complete Turonian Stage (Amon *et al.*, 1987). Similar radiolarian assemblages are also known from other Turonian sequences studied further eastward in West Siberia (Lipman, 1962; Kozlova and Gorbovets, 1966).

Bed 9 (45.8–47.0 m, core recovery 1.2 m): carbonate-free and slightly micaceous meager clay darker in color than the overlying bed. The rock shows thin lenticular intercalations of light gray silt. The lower boundary is fairly distinct.

The bed yielded scarce foraminifers *Gyroidinoides* sp. and *Cibicidoides apprima* (Wolosch.) associated with radiolarians *Orbiculiforma cachensis* Pess., *O. maxima* Pess., *O. multangula* Pess., *Pessagnobrachia clavata* (Squinabol), and *P. fabianii* (Squinabol). Two last forms (Plate II) are typical of lower Turonian strata in Italy (O'Dogherty, 1994) and Slovakia (Sykora *et al.*, 1997).

Paramonovo Formation

Bed 10 (47.0–49.8 m, core recovery 2.8 m): dark gray, slightly greenish silt coarse at the top and fine at the base; the admixture of clay and mica is irregularly distributed. The bed encloses thin (up to 3 mm) laminas of dark gray silty clay. This and underlying beds grade one into another.

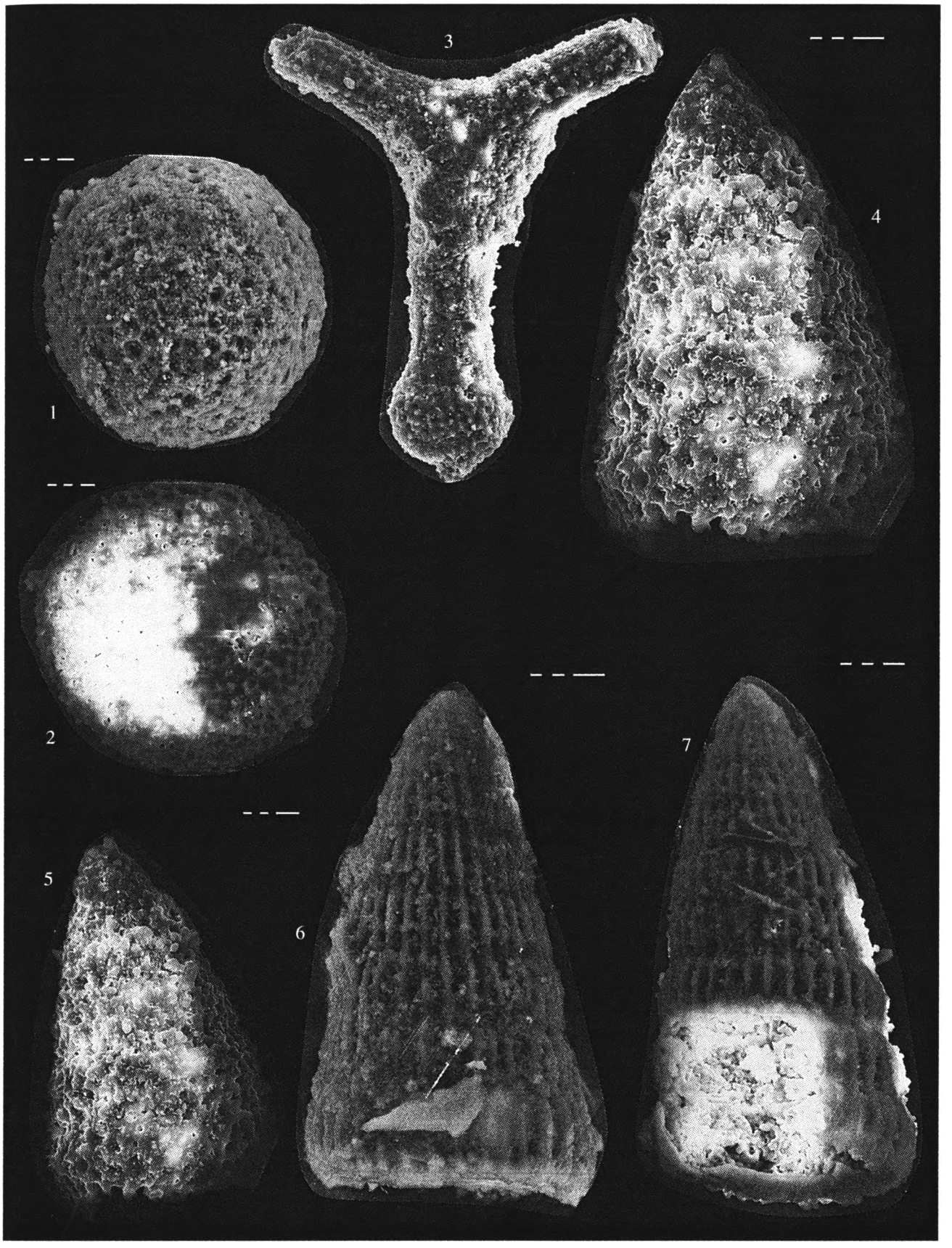
Bed 11 (49.8–54.0 m, core recovery 4.0 m): dark gray, bioturbated meager clay with the fine silty admixture. Narrow and branching fucoids are chaotically distributed and filled with light gray silt. Percentage of silt admixture decreases downward. Beds 10 and 11 correspond together to a single sedimentation rhythm. The lower boundary is distinct.

Bed 12 (54.0–56.0 m, core recovery 2.0 m): gray coarse quartz silt. The content of clayey admixture increases downward, where thin clay laminas are isolated. The laminae are slightly bioturbated and have some silty admixture. The transition to underlying rocks is gradual.

Bed 13 (56.0–61.0 m, core recovery 3.4 m): dark gray meager clay enclosing thin lentils and wedging interlayers of light gray silt near the top of the bed. Abundance of silty material increases downward.

Beginning from the depth level of 53.5 m (beds 11–13), rocks of the Paramonovo Formation yield the late Albian radiolarians of the *Porodiscus kavilkinensis–Crolanium cuneatum* assemblage (Kazintsova *et al.*, 1997). Despite the absence of *Crolanium* forms in the studied section, other species characteristic of the reference assemblage are present. These are *Porodiscus inflatus* Smirn. et Aliev, *P. kavilkinensis* Aliev, *Orbiculiforma multangula* Pess., *O. nevadaensis* Pess., *Sti-*

Plate II. Radiolarians from the Chernevo Formation (lower Turonian), Borehole 102, Chernevo site of the Yaroslavl region and Andreevskoe site of the Vladimir region: (1) *Cavidiscus fiskensis* Pessagno, ×300, Borehole 102, depth 38.0 m, lower Turonian; (2) *Orbiculiforma cachensis* Pessagno, ×300, Borehole 102, depth 43.0 m, lower Turonian; (3) *O. vacaensis* Pessagno, ×400, Andreevskoe site, depth 13.1 m, lower Turonian; (4) *Crucella cachensis* Pessagno, ×400, Andreevskoe site, depth 11.6 m, lower Turonian; (5) *Pessagnobrachia fabiani* (Squinabol), ×220, Borehole 102, depth 46.0 m, lower Turonian; (6) *Pessagnobrachia clavata* (Squinabol), ×250, Borehole 102, depth 46.0 m, lower Turonian.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chocapsa ferosia (Aliev), and *Dictyomitra konachkendensis* Aliev.

Andreevskoe Section (Vladimir Region)

This section recommended for the Chernevo Formation parastratotype is exposed 34 km away to the northwest from Vladimir. Exposures are located in a ravine open toward the Yakhroma River valley. Relative to the adjacent reference points, the site is situated 0.6 km to the east-southeast from the monument in the Village of Andreevskoe (Yur'ev-Pol'skii area) and 0.3 km to the west from the highway Vladimir–Yur'ev-Pol'skii. The following stratigraphic units (Fig. 6) are exposed here (from the base upward):

Paramonovo Formation

Bed 1: dark gray compact silty–micaceous clay with some signs of bioturbation and pyrite segregations. The bed is 1.0 m thick.

Scarce radiolarians from the bed are *Orbiculiforma multangula* Pess., *O. nevadaensis* Pess., *Stichocapsa ferosia* (Aliev), and *Theocampe cylindrica* Smirn. et Aliev. All species are characteristic of the Paramonovo Formation (Kazintsova *et al.*, 1997).

Chernevo Formation

Bed 2: dark gray to slightly greenish fine-grained quartz sand with an admixture of silty and clayey material. The bed lacks visible layering and shows spots of a lighter color. Phosphorite nodules (1–2 cm across) are dispersed throughout the bed being more concentrated near the base, where they are larger (up to 4 cm across) and associate with rolls of siderite clay and oxidized pyrite segregations. The bed is 0.9 m thick.

Bed 3: gray to light gray calcareous clay changing composition to the category of clayey marl. Being bioturbated, the basal portion of the bed (0.3 m thick) is rich in silty material and encloses small (to 1 cm across) figurate phosphorite nodules. Fucoids are to 1 mm in diameter and to 4 cm long. Rare lentils of pale silty material and fine-grained shell detritus appear in the upper part of the bed. Having gradual transition to underlying sediments, clay of the bed is eluviated and ferruginated near the top. Its upper boundary is very distinct. The bed is 3.0 m thick.

Both beds of the Chernevo Formation bear foraminifers. *Gaudryina tricarinata* (Orb.), *G. serrata* Franke, *Gaudryinopsis* aff. *filiformis* (Berth.), and *Marssonella* cf. *oxycona* Reuss are forms typical of sand. The foraminiferal assemblage from basal part of the clay bed is more diverse owing to appearance of small secretory forms. Among the latter, there are *Globorotalites turonicus* Kaever, *Eponides belorussiensis* Akim., *E. monterelensis* Marie, *E. turonicus* Lipn., *Gavelinella vesca* (N. Byk.), *Cibicides polyrraphes polyrraphes* (Reuss, and *C. lepidus* Plotn. usually characteristic of lower Turonian deposits. Agglutinating Turonian species *Arenobulimina minima* Vass. appear at this level together with *Gaudryinopsis filiformis* (Berth.) and *Ammobaculites fragmentarius agglutiniformis* Pod. typical of this stage in West Siberia. The complete composition of the assemblage is shown in Fig. 6. The upper portion of the clay bed yields rare shells of agglutinating foraminifers *Gaudryina arenosa* Akim., *G. tricarinata* (Orb.), *G. serrata* Franke, *Gaudryinopsis filiformis* Berth., *Ammobaculites fragmentarius agglutiniformis* Pod., and *Trochammina subbotinae* Zasp.

In addition to foraminifers, Bed 3 produced radiolarians *Crucella cachensis* Pess., *Orbiculiforma multa* (Kozl.), *O. impressa* (Lipm., *O. maxima* Pess., *O. vacaensis* Pess., and *Cromyodruppa concentrica* Lipm. These species are of a wide stratigraphic range, but they associate with *Cenosphaera magna* Grig., *Stichocapsa pyramongata* (Grig.), and *Dictyomitra napaensis* Pess. (Plate III) whose distribution is limited by the Turonian Stage.

Zagorsk Formation

Bed 4: gray fine-grained glauconite sandstone with dark gray patches and uniformly dispersed gravel and pebbles. The rock matrix consists of opal. Glauconite is frequently oxidized and replaced with Fe-hydroxides. The bed progressively grades into the overlying layer, but its lower boundary is distinct. The bed is 0.3 m thick.

Bed 5: light gray tripoli grading into gray opoka. The latter is irregularly silicified (dark gray chaotic patches) and shows signs of bioturbation. The bed is 0.5 m thick.

Bed 6: light gray tripoli (greenish brown when wet). In places, the rock is vaguely bedded because of alternation of lighter and darker laminas. At the top, there are burrows filled with sandy material derived from the overlying bed (an immature hardground), and the upper boundary is very distinct. The bed is 4.2 m thick.

Tripoli and opoka samples from beds 5 and 6 yield radiolarians, whose assemblage includes *Orbiculiforma multa* (Kozl.), *Cromyodruppa concentrica*

Plate III. Radiolarians from the Chernevo Formation (lower Turonian), Borehole 102, Chernevo site of the Yaroslavl region and Andreevskoe site of the Vladimir region: (1) *Cenosphaera magna* Grigorieva, ×300, Andreevskoe site, depth 12.1 m, lower Turonian; (2) *Orbiculiforma maxima* Pessagno, ×270, Andreevskoe site, depth 11.6 m, lower Turonian; (3) *Patulibrachium irregulare* (Squinabol), ×300, Borehole 102, depth 25.0 m, lower Turonian; (4, 5) *Stichocapsa pyramongata* (Grig.), ×600 and ×400, Andreevskoe site, depth 11.1 m, lower Turonian; (6, 7) *Dictyomitra multicostata* (Squinabol), ×600 and ×300, Borehole 102, depth 43.0 m, lower Turonian.

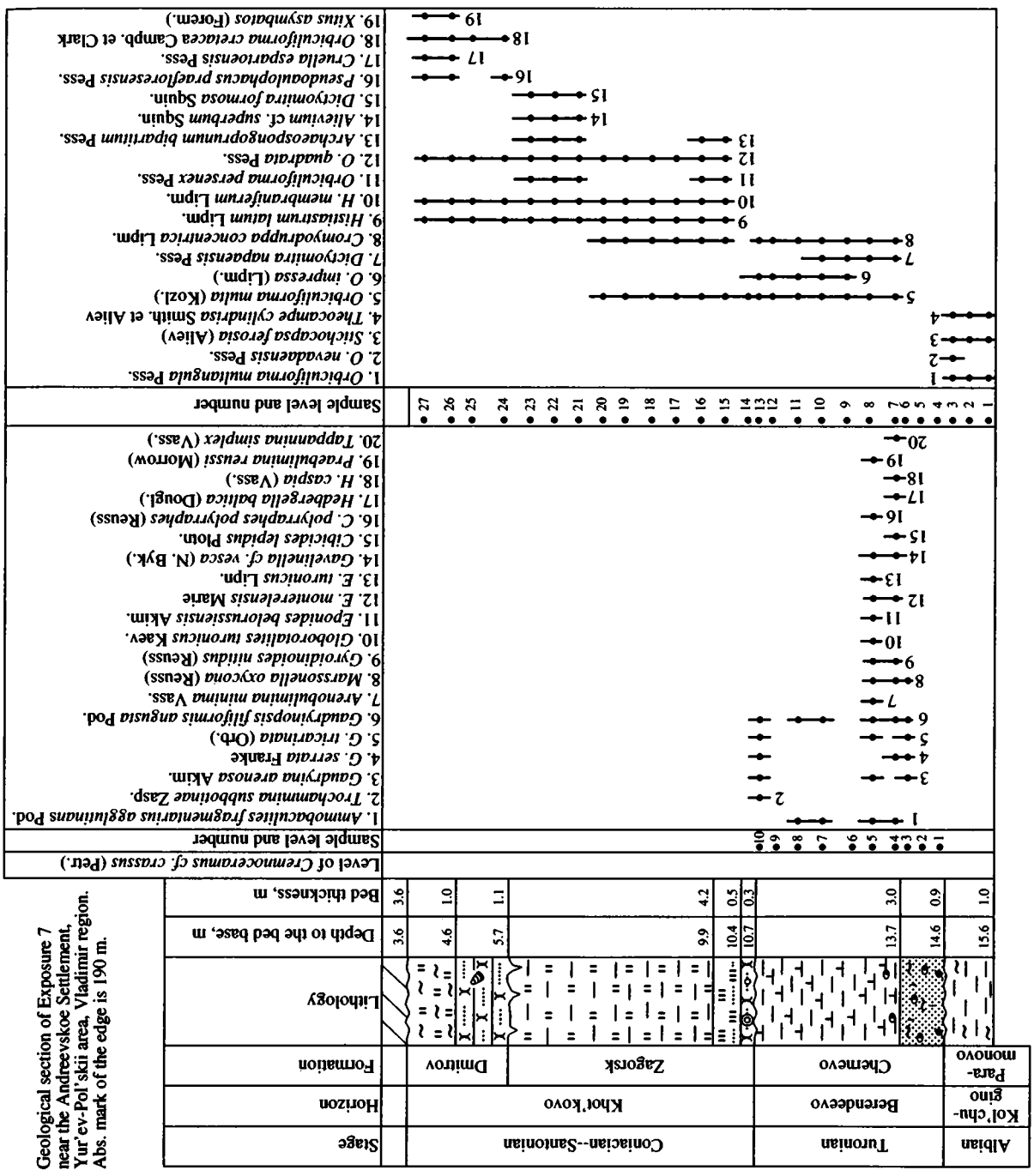

Fig. 6. Distribution of fossils in Cretaceous sediments, Exposure 7 near the village of Andreevskoe (symbols as in Fig. 2).

Lipm., *Histiastrum latum* Lipm., and *H. membraniferum* Lipm. These forms of a wide stratigraphic range coexist with *Orbiculiforma persenex* Pess., *O. quadrata* Pess., *Archaeospongoprunum bipartitum* Pess., and *Dictyomitra formosa* Squin. The last group is thought to be indicative of basal beds of the Zagorsk Formation, the only unit in the Moscow syneclise, where we detected *Alievium superbum* Squin. typical of the Turonian-Coniacian biozones in the Voronezh anteclise, Crimea, Ciscaucasia, Caucasus, Mediterranean region, and California.

Dmitrov Formation

Bed 7: gray to greenish fine-grained glauconite sandstone. Siliceous, unevenly distributed matrix is responsible for patchy appearance of the rock grading progressively into the overlying sediment. In our collection of inoceram casts, Walaszyk identified *Cremnoceramus crassus* (Petr.) typical, in his opinion, of the basal middle Coniacian. The Bed is 1.1 m thick.

Bed 8: silt grading into tripoli, light gray to greenish brown when wet. The bed is 1.0 m thick.

In addition to radiolarian species inherited from underlying strata, beds 7 and 8 yield radiolarians of a wide stratigraphic range: *Pseudoaulophacus praefloresensis* Pess., *Orbiculiforma cretacea* (Camp. et Cl.), *Crucella espartoensis* Pess., and *Xitus asymbatos* (Forem.). These forms contribute almost nothing to the age determination of the Dmitrov Formation, and the first species solely suggests that the formation is not younger than the early Campanian.

The described sequence is overlain by the Quaternary loam.

DISCUSSION

Chernevo Formation and Its Position in the Sequence

The described stratotypes of the Chernevo Formation overlie the upper Albian Paramonovo Clay with the stratigraphic hiatus. Only in the extreme west (Borehole 130 near the village of Lobkovo, Aleksandrov area of the Vladimir region), deposits of this formation rest on Cenomanian sands of the Yakhroma Formation of the Varavino Group. Actually, this is the sole section, where the Chernevo Formation overlies Cenomanian deposits, unless we do not consider as Cenomanian the sands of problematic age, which were penetrated in the depth interval of 33.4–36.2 m by Borehole 163 near the village of Savel'evo of the Kol'chugino area. Attributed to the Cenomanian in the course of geological survey, these sands are almost free of carbonates and bear glauconite. In our opinion, both of these features suggest their similarity with the basal sand bed of the Chernevo Formation that was studied in exposures of the Andreevskoe site. Barren sand and clay beds penetrated by Borehole 200 near the village of Bakino are conditionally attributed to the Chernevo Formation, and their position above the Cenomanian Yakhroma Formation, as it is shown in profile I–I (Fig. 2), appears to be doubtful.

The Chernevo Formation is composed almost everywhere of calcareous clay locally grading into marl facies and includes basal beds of sand and silt. Siliceous rocks, like tripoli, opoka, silica-rich clay, and sandstone with siliceous matrix, are missing from its composition. All these rocks are typical of the Khot'kovo Horizon and overlie the Chernevo Formation with obvious signs of erosion, truncating the rocks of the latter in western and eastern directions so that they completely disappear from the sections of Upper Cretaceous deposits.

The Turonian age of the Chernevo Formation is established on the basis of belemnite species *Goniocamax intermedius* (Arkh.) that have been found in clay beds of borehole sections 6 and 92 recovered near the villages of Spas (Pereyaslavl'-Zalesskii district, Yaroslavl' region) and Makarovo (Kol'chugino district, Vladimir region). *Inoceramus lamarcki* Park. that was detected by A.M. Viktorov in 1934 near the village of Denis'evo, not far away from the Andreevskoe section, suggests the same conclusion. Affiliation of the Chernevo Formation with the Turonian Stage is also evident from foraminiferal and radiolarian assemblages occurring in respective sediments.

Turonian Microfauna

Foraminifers. The foraminiferal assemblage from clay beds of the Chernevo Formation (Plate I) is represented, on the one hand, by a series of species well known in the Ukrainian and Baltic syneclises. These are *Spiroplectammina cuneata* Vass., *S. praelonga* (Reuss), *Gaudryina angustata* Akim., *G. arenosa* Akim., *G. serrata* Franke, *G. tricarinata* (Orb.), *Arenobulimina improcera* Wolosch., *A. minima* Vass., *A. presli* (Reuss), *Eggerellina intermedia* (Reuss), *Marssonella oxycona* Reuss, *Valvulineria lenticula* (Reuss), *Gyroidinoides nitidus* (Reuss), *Globorotalites multiseptus* Brotz., *G. turonicus* Kaev., *Eponides belorussiensis* Akim., *E. monterelensis* Marie, *E. turonicus* Lipn., *Gavelinella nana* Akim., *Brotzenella belorussica* (Akim.), *B. berthelini* (Kell.), *Lingulogavelinella globosa* (Brotz.), *G. vesca* (N. Byk.), *Cibicides lepidus* Plotn., *C. minusculus* Akim., *C. polyrraphes polyrraphes* (Reuss), *Hedbergella caspia* Vass., *H. baltica* Dougl., *H.* aff. *infracretacea* (Glaessn.), *Praeglobotruncana umbricata* (Mornod.), *Globotruncana globigerinoides* (Brotz.), *Praebulimina reussi* (Morrow), *Reussella turonica* Akim., *Sitella gracilis* Vass., *Eouvigerina regularis* (Kell.), *Tappannina eouvigeriniformis* (Kell.), *T. simplex* (Vass.), and *Heterohelix globulosa* (Her.). All these species in total define the early Turonian age of their host deposits. On the other hand, the assemblage includes forms typical of the lower Turonian *Gaudryinopsis filiformis angusta* Zone distinguished in West Siberia. In addition to the mentioned

Radiolarians from Turonian deposits of the Russian platform, Trans-Urals, and West Siberian plate

Radiolarian species	Russian platform	Trans-Urals	West Siberian plate
	Lipman, 1952; Vishnevskaya, 1987; and this work	Grigor'eva, 1975; Amon, 1985, 1987, 1994	Lipman, 1962; Kozlova and Gorbovets, 1966
Cenosphaera minor Lipman	+	+	
C. magna Grigorjeva	+	+	
Conosphaera cf. *mammilata* Lipman	+		+
C. ex gr. *clivosa* Lipman		+	+
Cavidiscus fiskensis Pessagno	+		
Elliphoxiphus cf. *asper* Rüst		+	
Cromyodruppa concentrica Lipman	+		+
Trochodiscus cf. *spiniger* Lipman		+	
Triacticus cf. *triacuminatus* Lipman		+	
Porodiscus vulgaris Lipman	+	+	+
Ommatodiscus modilis Kozlova	+	+	+
Prunobrachium articulatum Lipman			+
Stylodictya placentalis Lipman	+		
S. delicatula Lipman	+	+	
Amphibrachium spongiosum Lipman	+		+
A. concentricum Lipman			+
Histiastrum latum Lipman	+	+	
H. tetracantum Lipman	+	+	
Rhopalastrum attenuatum Lipman	+		
Pessagnobrachia clavata (Squinabol)	+		
P. fabianii (Squinabol)	+		
Patulibracchium irregulare (Squinabol)	+		
Spongodiscus volgensis Lipman	+	+	+
S. maximus Lipman	+		
Spongotripus aculeatus Lipman	+		
Stylotrochus dolichacanthus Lipman	+	+	
Orbiculiforma cacchensis Pessagno	+		
O. impressa (Lipman)	+	+	+
O. maxima Pessagno	+		
O. vacaensis Pessagno	+		
O. multa (Kozlova)	+	+	+
Crucella aster (Lipman)		+	+
C. cachensis Pessagno	+		
C. messinae Pessagno	+		
Halesium sp. aff. *sexangulum* Pessagno	+		
Stichocapsa pyramidata (Grigorjeva)	+	+	+
Dictyomitra formosa Squin.	+		
Dictyomitra multicostata Zittel	+		+
D. napaensis Pess.	+		
D. striata Lipman	+		
Amphipyndax uralicus (Gorbovetz)		+	

index species, this zonal assemblage includes *Textularia anceps* Reuss, *Trochammina subbotinae* Zasp., *Nodosaria orthopleura* (Reuss), *N. tubifera* Reuss, *Reophax texanus* Cush. et Wat., *Ammobaculites tyrrelli agglutinoides* Dain, *Bagginoides quadrilobus* (Mello), *Ataxophragmium initiale* Wolosch., and *Haplophragmoides rota sibiricus* Zasp.

In the most complete sections, the formation under consideration is overlain by the sandy Potanino Beds, where the foraminiferal assemblage includes a series of newcomers. These are *Arenobulimina* aff. *elevata* (Orb.), *Ataxophragmium nautiloides* Brotz., *Stensioeina* aff. *emscherica* Bar., *S. praeexsculpta laevigata* Akim., *Osangularia whitei* (Brotz.), *Gavelinella ammonoides* (Reuss), *G.* cf. *giedroyci* Grig., *G.* cf. *kelleri* (Mjatl.), *G.* cf. *praeinfrasantonica* (Mjatl.), *Cibicidoides praeeriksdalensis* (Vass.), *Hedbergella* aff. *agalarovae* Vass., and *Reussella kelleri* Vass. This group indicates the early Coniacian age of the beds.

Radiolarians. Radiolarians from clay beds of the Chernevo Formation stratotype turned out to be more diverse and better preserved than in other coeval assemblages from the Russian platform. It should be mentioned that Lipman (1952) was first to describe an assemblage of presumably Turonian radiolarians from green clay beds underlying Santonian deposits in the Kuznetsk area of the Penza region. This assemblage consisting of *Cenosphaera minor* Lipm., *Carposphaera* sp., *Stylodictya placentalis* Lipm., *St. delicatula* Lipm., *Rhopalastrum attenuatum* Lipm., *Spongodiscus volgensis* Lipm., *Spongotripus aculeatus* Lipm., *Stylotrochus dolichacanthus* Lipm., and *Orbiculiforma impressa* (Lipm.) was later termed as the *Stylodictya delicatula–Spongotripus aculeatus* radiolarian fauna of the Turonian Stage (Vishnevskaya *et al.*, 1990). In 1964, Smirnova (see in Lavrova, 1971) detected another radiolarian assemblage of a similar composition in the Turonian deposits of the Moscow syneclise.

Afterward, Vishnevskaya (1987) reported on Turonian radiolarians from the Ukrainian syneclise (Bryansk, Kaluga, and Orlov regions). She defined them as the *Spongotripus aculeatus–Stichocapsa pyramongata* assemblage consisting, in addition to index species, of *Crucella messinae* Pess, *Halesium* sp. aff. *H. sexangulum* Pess, *Dictyomitra formosa* Squin., and *D. striata* Lipm. In this work, it was noted for the first time that the Turonian assemblage from the Moscow syneclise includes species characteristic of coeval deposits in California.

Kazintsova, who studied radiolarians from Turonian deposits recovered by Borehole 105 (Ivorovo site in the Yur'ev-Pol'skii area of the Vladimir region), defined here *Crucella cachensis* Pess., *Orbiculiforma maxima* Pess., *Pseudoaulophacus* sp., *Alievium* sp., and *Dictyomitra napaensis* Pess.

The Chernevo Formation stratotype yielded an assemblage consisting of 20 radiolarian species, mostly of spumellarians except for two nassellarian forms. Kazintsova suggested to term the discovered community as the *Spongodiscus volgensis–Stichocapsa pyramongata* assemblage. Spherical skeletons of *Cenosphaera* along with large and thick spongy forms of *Spongodiscus* and *Orbiculiforma* characterize, together with smooth-walled five- to seven-chamber *Stichocapsa* and *Dictyomitra*, the general composition of the assemblage. *Cenosphaera minor* Lipm., *Porodiscus vulgaris* Lipm., *Spongodiscus volgensis* Lipm., *Spongotripus aculeatis* Lipm., *Orbiculiforma impressa* (Lipm.), and *Stichocapsa pyramongata* (Grig.) are most characteristic species of the assemblage. In addition to smooth-walled spherical spumellarians of the genus *Cenosphaera*, we encountered some *Conosphaera* forms decorated with tubercles. Discoid forms belong to families Porodiscidae and Spongodiscidae, the first of which consists of five genera though represented by few specimens, namely by round or almost round disks (*Porodiscus*, *Stylodictya*, and scarce *Ommatodiscus* species) and also by square (*Histiastrum*) and axially elongated forms with two simple processes at both ends of the axis (*Amphibrachium* species also scarce). Representatives of Spongodiscidae are abundant; thick discoid *Spongodiscus volgensis* Lipm. and *Orbiculiforma impressa* (Lipm.) are dominant among them, whereas *Stylotrochus* and *Spongotripus* species occur as single specimens. Nasselarians are also abundant, but only *Stichocapsa pyramongata* (Grig.), *Dictyomitra napaensis* Pess., and *D. multicostata* Zittel represent this group.

In its taxonomic composition, the *Spongodiscus volgensis–Stichocapsa pyramongata* assemblage is similar to radiolarian assemblages from the Turonian Kuznetsov Formation of the Trans-Urals (Grigor'eva, 1975; Amon, 1985) and West Siberia (Lipman, 1962; Kozlova *et al.*, 1966) being much more diverse than the early Turonian assemblage from Poland (Gorka, 1995). Like in the Chernevo Formation, Turonian radiolarians from above regions are mostly represented by spongy spherical and discoid forms, whereas the associated nasselarians belong to genera *Stichocapsa* and *Dictyomitra*. Species in common for all regions (Plate I) are *Porodiscus vulgaris* Lipm., *Ommatodiscus mobilis* Kozl., *Spongodiscus volgensis* Lipm., *Orbiculiforma multa* (Kozl.), *O. impressa* (Lipm.), and *Stichocapsa pyramongata* (Grig.). These species represent up to 55% of Turonian assemblages known from the Russian platform, West Siberia, and Trans-Urals, although the assemblages have their own individual feature in each region. For instance, *Elliphoxiphus*, *Trochodiscus*, and *Triactiscus* species typical of assemblages from the Trans-Urals do not occur in the Russian and West Siberian platforms. At the same time, Turonian radiolarians representing genera *Stylodictya*, *Amphibrachium*, *Rhopalastrum*, *Spongotripus*, and *Halesium* in European Russia are unknown immediately behind the Urals, whereas *Stylodictya*, *Histiastrum*, *Rhopalastrum*,

Spongotripus, and *Stylotrochus* forms are missing from radiolarian assemblages of West Siberian plate.

CONCLUSION

(1) The early Turonian age of the Chernevo Formation is substantiated on the basis of radiolarian and foraminiferal microfaunas.

(2) Taking into consideration the well-known standard curve of eustatic sea-level fluctuations (Haq *et al.*, 1987), we can see that the respective events have been recorded in the Moscow region not only during the Aptian–Albian, as noted by Alekseev *et al.* (1996), but also during the Turonian time.

(3) Despite the particular provincial features of radiolarian and foraminiferal assemblages, we may state that sedimentary basins in the Russian and West Siberian platforms definitely had connections during the Turonian Age. The Urals did not represent a barrier for these connections, because the sea level rise was global at that time (Haq *et al.*, 1987). Data obtained recently suggest that the mountain chain of the Urals was high during the Late Permian and Triassic. However, geological records suggesting that the Urals separated the above sedimentary basins since the Late Jurassic until Paleogene remain unknown. It is not inconceivable that the Uralian belt subsided during this period under the sea level, partially at least, and rose again as the chain of mountains in the post-Neogene time, but this is a topic for another paper.

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Miocene Stratigraphy and Geological History of the Crimean Plain Region

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Abstract—Local Miocene Stratigraphic Schemes for different plain areas in the Crimea are substantiated and correlated with the Crimea–Caucasus regional scale validated in adjacent territories. New data are used to elucidate the Miocene geological history of the Crimean Plain.

Key words: stratigraphy, geological history, Crimean Plain, Miocene, mollusks, foraminifers.

The last fundamental book on the Crimean geology was issued over 30 years ago (*Geologiya SSSR...*, 1969). In subsequent period, the intermediate- to large-scale maps were compiled for the whole plain area of the peninsula in order to be used as a basis for the mineral prospecting.

Fossils from Miocene deposits in the territory under consideration are inadequately characterized even now. Available episodic publications mostly report on molluscan faunas associated with less abundant foraminifers and ostracodes (*Atlas...*, 1959; Konenkova, 1991; Pishchikova, 1992).

New paleontological materials obtained recently (figure) are used as a substantiation of several local stratigraphic schemes correlated with the Regional Stratigraphic Scale of the Crimea, Caucasus, and adjacent regions (see the table). These results and additional information from the eastern Paratethys (Il'ina *et al.*, 1976; Il'ina, 1980, 1995; Nevesskaya *et al.*, 1984, 1986; Goncharova, 1989; Goncharova and Shcherba, 1997; Goncharova and Il'ina, 1997; Popov *et al.*, 1993; Popov, 1996) elucidate the Miocene geological history of the Crimean Plain.

The oldest Miocene rocks in the study region are those of the upper Kerleut Subformation distinguished in the upper part of the Regional Caucasian Stage (Nosovskii and Bogdanovich, 1980). In the Indol and northern areas, deposits of this age are represented by greenish to dark gray silty clay intercalated with siltstone interbeds and bearing radiolarians, sponge spicules, and fish remains. In these clay beds (Borehole 42, 3 km to the west of the village of Zernovoe), Konenkova (1991) detected an assemblage of foraminifers *Spiroplectammina caucasica* Djan., *Uvigerinella californica* Cushm., *Caucasina schischkinskyae* (Saml.), *Bolivina* aff. *punctata* Orb., and *Melonis* sp. The beds are 20 m thick in the borehole section, but their thickness grows up to 400 m in the Indol depression. The subformation is thought to be correlative with the Gornostaevka Formation of the northern Black Sea region, where it is included into the Regional Caucasian Stage (Nosovskii and Bogdanovich, 1980).

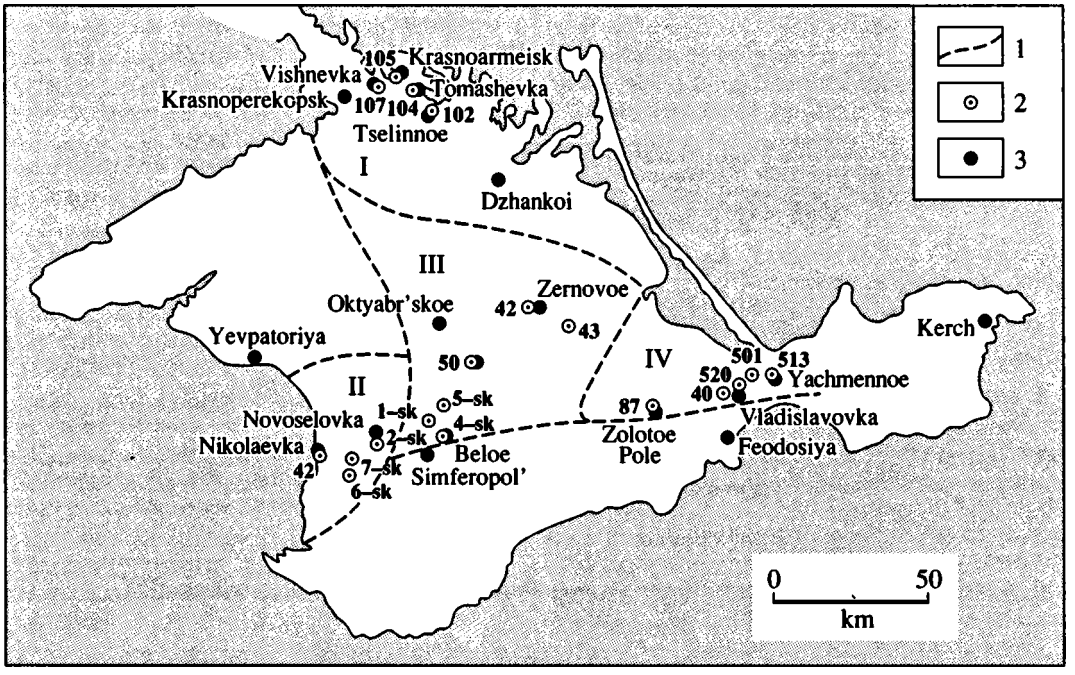
Regional Sakaraulian Stage, Arbatskaya Formation. The unit was originally distinguished by V.F. Kozyreva as the *Bathysiphonus* Beds. According to regulations of the Stratigraphic Code, it was renamed as the Arbatskaya Formation on the basis of later data of L.M. Golubnichaya (see in *Geologiya SSSR...*, 1969).

In the Indol area (Borehole 87, 1.5 km to the northwest of the village of Zolotoe Pole), the formation corresponds to a sequence (52 m thick) of gray carbonate-free silty clay presumably resting on the upper Kerleut Beds and discordantly underlying the Sartagan Beds of the Konkian regional stage. The clay yielded foraminifers *Neobulimina elongata* (Orb.), *Bulimina* sp., *Caucasina schischkinskyae* (Saml.) *Uvigerinella parkeri breviformis* Papp et Turn., *Globulina pupoides* (Orb.), *Elphidium rugosum* (Orb.), and *Ammonia* sp. (determinations of I.V. Venglinskii).

Borehole 40 penetrated similar deposits to the northwest of the village of Vladislavovka at the Mokryi Indol River.

According to drilling data, the formation is as thick as 1200 m in total. Fossils from this unit suggest its correlation with the Chernobaevka Formation in the northern Black Sea region and with the Ol'ginskaya Formation in the western Ciscaucasia.

Regional Kozachurian Stage, Korolevo Formation. In the Indol area, the formation is up to 300 m thick and consists of dark gray, slightly brownish silty clay bearing scarce foraminifers *Saccammina zuramakensis*



Borehole sites in the Crimean Plain region: (1) boundaries between northern (I), Al'ma (II), central (III), and eastern (IV) areas; (2) borehole; (3) settlement.

Bogd. Deposits of this formation are coeval to the Karga Beds of the northern Black Sea Region (Barg, 1993).

Regional Tarkhanian Stage. New paleontological data show that deposits of this level are widespread in the Crimean Plain. The most complete sections of the stage are those discovered in the Akmanaiskii Isthmus and Al'ma depression. These two areas experienced an intense subsidence during the early Miocene (Chekunov *et al.*, 1976).

The sedimentary succession near the village of Yachmennoe, where Borehole 513 penetrated Tshokrakian, Tarkhanian, and upper Maikop beds, can be considered as the reference section for the eastern part of the Crimean Plain. The Tarkhanian interval corresponds here to the 26-m-thick sequence of greenish gray silty clay divided into Kamyshlak, Tarkhan, and Yurakov beds described recently in detail (Barg and Ivanova, 1997). Planktonic and benthic foraminifers *Globigerina bollii* Cita et Premoli Silva, *Globigerinoides* cf. *trilobus* (Reuss), *Spiroloculina bicarinata* Djan., and *Glabratella tarchanensis* (Djan.) are most significant stratigraphic markers of the stage.

In the Al'ma depression, Tarkhanian deposits were penetrated by three boreholes: 6-sk, 2 km to southwest of Prudovoe; 7-sk, 3 km to the south of Kol'chugino (Ivanova *et al.*, 1998); and 42 near Nikolaevka (Barg *et al.*, 1992). The deposits overlie the Maikop clay beds with siliceous microfossils and underlie the Tshokrakian shallow-water sediments bearing characteristic mollusks and foraminifers. The stage is subdivided in this area into the Tarkhan (s. str.) and Yurakov beds.

The Tarkhan Beds of greenish gray clay yield the following fossil groups: mollusks *Chlamys* aff. *macrotis* (Sow.); planktonic foraminifers *Globigerina* aff. *tarchanensis* Subb. et Chutz., *G. bradyi* Wissner, *G. officinalis* Subb., *Globoquadrina* aff. *larmeui* Akers, *Turborotalia mayeri* (Cushm. Et Ell.), and *Globigerinoides bisphaericus* Todd.; benthic foraminifers represented by abundant polymophinids, rotaliids, and nonionids; diverse nannoplankton characteristic of the *Helicosphaera ampliaperta* Zone NN 4 (Bogdanovich, 1998); worm tubes (*Pectinariopsis*); sponge spicules; echinoid spines; ostracodes; and gyrogonites of charophytes. The beds are 4.5 m thick.

Fossils from the Yurakov Beds of greenish gray and light green clays are also represented by several groups: pteropods *Spiratella andrussovi tshokrakensis* (Zhizh.); planktonic foraminifers *Globigerina* aff. *bollii* Cita et Premoli Silva, *G. bulloides* Orb., *G. praebulloides* Blow, *Globigerinoides trilobus* (Reuss), *G. bisphaericus* Todd, *Globoquadrina dehiscens* (Chap., Parr et Coll.), and *Turborotalia mayeri* (Cushm. et Ell.); benthic foraminifers (polymorphinids, lagenids, rotaliids, nonionids, anomalinids, bolivinitids, and others); nannoplankton of the *Helicosphaera ampliaperta* Zone NN 4 (Bogdanovich, 1998); ostracodes; worm tubes (*Pectinariopsis*); sponge spicules, bryozoans; fish teeth; and gyrogonites of charophytes. The beds are 0.6–5.0 m thick.

In both the Al'ma depression and Akmanaiskii Isthmus, the uppermost portion of Yurakov Beds shows a sharp diversity decrease in microfauna consisting of euryhaline foraminifer species.

Around the Gulf of Sivash, deposits of the regional Tarkhanian Stage were detected on the basis of indicative ostracod remains (Pishchikova, 1992).

The analyzed Tarkhanian fossils from the Crimean Plain, especially the planktonic foraminifers of a high stratigraphic potential and *Chlamys* aff. *macrotis* (Sow.) identified here for the first time, show that their host deposits can be correlated beyond doubt with the following stratigraphic subdivisions: the Carpathian Stage of western and central Europe (Roger, 1939); the Balichi Formation of the Ciscarpathian region; the Nagoryany, Berezhany, and Tomakovka beds of Ukraine; the Tarkhanian Stage in the Kerchenskii Peninsula and coeval strata in the northern Caucasus and southern Ciscaucasia (*Stratigrafiya SSSR...*, 1986; Barg, 1993)

Regional Tshokrakian Stage. Deposits of this level occur in many plain areas of Crimea. They are frequently barren or yield an impoverished molluscan fauna (shell impressions and casts) and associated scarce foraminifers. In many cases, the Tshokrakian interval is outlined tentatively on the basis of its lithological features and stratigraphic position in the sections.

Near the village of Yachmennoe (Borehole 513), the Tshokrakian Stage spans the light and greenish gray beds of oolitic limestone that yielded mollusks *Bittium digitatum* Zhizh., *Gibbula kertschenensis* Uspensk., *Spaniodontella intermedia* Bajar., and *Cerastoderma* sp. in association with foraminifers *Sigmoilinita* cf. *tschokrakensis* (Gerke), *S. haidingerii aculeata* (Bogd.), *Ammonia beccarii* (L.) and *Elphidium rugosum atschiensis* Suzin, and also with ostracodes, bryozoans, and fish otoliths. These deposits resting on the Yurakov Beds of the Tarkhanian Stage are overlain by the Karaganian strata. The described unit is 3.0 m thick.

At the other site near the village of Vladislavovka (Borehole 40), the Tshokrakian interval corresponds to green beds of silty clay intercalated with rare sandstone interlayers. *Ervilia praepodolica* Andruss., *Acanthocardia centumpania* (Andruss.), and *Gibbula pictiformis* (Andruss.) represent mollusks occurring in these 10.0-m-thick beds. Konenkova (1991) detected here foraminifers *Elphidium* ex gr. *rugosum atschiensis* Suzin, *E. cubanicum* Krash., *E. kudakoense* Bogd., *Florilus boueanus* (Orb.), *Discorbis figuratus* Konen., *Ammonia mjatliukae* (Putrja), and *Nonion granosus parvus* Bogd.

In the northwestern slope of the Simferopol' Uplift (40 km to the east of the village of Oktyabr'skoe), Borehole 42 penetrated a sequence of fine-grained sandy beds enclosing interlayers of limestone and greenish gray clay. Rare Tshokrakian mollusks *Spaniodontella intermedia* (Andruss.), *Ervilia praepodolica* Andruss., and *Acanthocardia impar* (Zhizh.) are detected in this 18-m-thick sequence.

Near the Gulf of Sivash, Tshokrakian deposits were known in the Dzhankoi area (Pishchikova, 1992). We distinguished them also near the village of Vishnevka (Borehole 107), where they correspond to beds (3.8 m thick) of gray fine-grained detrital limestone. The beds yield foraminifers *Quinqueloculina* ex gr. *consobrina* (Orb.), *Flintina georgii* Bogd., *Nonion granosus parvus* Bogd., *N. bogdanowiczi* Volosh., *Protelphidium subgranosum* (Egger), *Ammonia galiciana* (Putrja), *Elphidium* sp., and *Fissurina* sp.

Borehole 6-sk penetrated the Tshokrakian sediments of the Al'ma depression at the site located 2 km to the southwest of the village of Prudovoe. In this case, sediments of the stage are represented by beds of light gray pelitomorphic limestone and sandstone. Mollusks *Ervilia* ex gr. *praepodolica* Andruss., *Gibbula pictiformis* (Andruss.), and *Acanthocardia* ex gr. *pseudomulticostata* Zhizh. associate in the beds with scarce sponge spicules, *Spirorbis* forms, and ostracodes. The beds are 6.0 m thick.

Molluscan and foraminifer species listed above existed in brackish-water basins of the Euxinian type formerly widespread in the Crimea–Caucasus region. They characterize an upper part of the Tshokrakian Stage that presumably corresponds to the range of the Bryk Beds (Goncharova, 1989).

Regional Karaganian Stage. Deposits of this stage are detected almost everywhere in the study region, where they transgressively overlie the Maikop, Tarkhanian, and Tshokrakian strata. There is a noticeable difference in lithology between western and eastern distribution areas of Karaganian sediments: clayey facies predominate within the southern slope of the Indol–Kuban' depression, and carbonate deposits are characteristic of the Al'ma depression and Simferopol' Uplift.

In the Akmanaiskii Isthmus near the village of Yachmennoe, Borehole 513 recovered alternating beds of light gray sandstone, green and dark gray to black silty clay, and greenish to green siltstone. The beds are up to 18 m thick. Their fossils are represented by abundant *Spaniodontella pulchella* Baily, scarce foraminifers *Nonion bogdanowiczi* Volosh. and *Quinqueloculina consobrina nitens* (Reuss), and also by sponge spicules and ostracodes. In the Indol area, Karaganian beds of sand and sandstone enclose clay interlayers bearing *Spaniodontella pulchella* Baily. In this area, they vary in thickness from 5 to 30 m.

In the Al'ma depression and Simferopol' Uplift (boreholes 1-sk, 4-sk, 5-sk, and 6-sk near the villages of Solnechnoe, Beloe, Chaikino, and Prudovoe), the Karaganian interval is composed of light gray recrystallized organogenic-detrital to oolitic limestones. *Spaniodontella pulchella* Baily, *Mohrensternia grandis* Andruss., and *M. barboti* Andruss. characterize molluscan fauna of these rocks. Associated foraminifers are

Stratigraphic scheme for the Miocene deposits in the Crimean Plain

Standard stratigraphic scale				Regional stratigraphic subdivisions				Local stratigraphic schemes	
Er-ath-em	Sys-tem	Se-ries	Sub-series	Regional stage	Regional substage	Formation, subformation	Beds	Northern areas	
Cenozoic	Neogene	Miocene	upper	Pontian	Bosphorian				
					Portafer				
					Novorossiisk		Odessa	Odessa	
							Yevpatoriya	Yevpatoriya	
				Maeotian	Akmanai		*Congeria*	*Congeria*	
					Bagerovo		*Dosinia*	*Dosinia*	
				Sarmatian	Kherson			Kherson	
					Bessarabian		Dnepropetrovsk	Dnepropetrovsk	
							Vasil'evka	Vasil'evka	
							Novomoskovsk	I	Novomoskovsk
					Volhynian		Zbruch		Zbruch
							Kuzhora		Kuzhora
			middle	Konkian			Veselyanka	Veselyanka	
							Sartagan	Sartagan	
							Ervilia–Pholadidae	*Ervilia*–Pholadidae	
				Karaganian				Karagan	
				Tshokrakian	upper		Bryk	Bryk	
					lower		Zyuk		
			lower	Tarkhanian			Yurakov	Yurakov	
							Tarkhan *s. str.*		
							Kamyshlak		
				Kozachurian		Korolevo			
				Sakaraulian		Arabatskaya			
				Caucasian		upper Kerleut			

(I) Krasnobaevka Formation.

Quinqueloculina ex gr. *consobrina* (Orb.), *Q.* ex gr. *reussi* (Bogd.), *Discorbis figuratus* Konen., *Ammonia* ex gr. *beccarii* (L.), *A. pseudobeccarii* (Putrja), *Protelphidium martkobi* (Bogd.), *P. subgranosum* (Egger), *Elphidium* aff. *rugosum* (Orb.), *E.* cf. *ligatum* Krash., *Nonion bogdanowiczi* Volosh., and *Nonionella* cf. *karaganica* Krash. Single bryozoans and *Spirorbis* forms represent other faunal groups. The beds under consideration are 10–22 m thick.

Near the Gulf of Sivash (in the vicinity of Krasnoperekopsk), Karaganian beds of sand, clay, and limestone bear mollusks *Spaniodontella pulchella* Baily and *Mohrensternia grandis* Andruss.

Owing to their molluscan and foraminiferal faunas, the Karaganian beds are easily detectable in geological sections. They are well correlated all over the Crimea–Caucasus region.

Stratigraphic schemes for adjacent regions				
Al'ma areas	Central areas	Indol areas	Kerch Peninsula	Northern Black Sea region
			Bosphorian	
			Portafer	
Odessa	Odessa	Novorossiisk	Novorossiisk	Novorossiisk
Yevpatoriya	Yevpatoriya			
Congeria	*Congeria*	*Congeria*	*Congeria*	*Congeria*
Dosinia	*Dosinia*	*Dosinia*	*Dosinia*	*Dosinia*
Kherson	Kherson	Kherson	Kherson	Kherson
Dnepropetrovsk	Dnepropetrovsk	Dnepropetrovsk	Dnepropetrovsk	Dnepropetrovsk
Vasil'evka	Vasil'evka	Vasil'evka	Vasil'evka	Vasil'evka
I Novomoskovsk	I Novomoskovsk	I Novomoskovsk	Novomoskovsk	Novomoskovsk
Zbruch	Zbruch	Zbruch	*Mactra*	Zbruch
Kuzhora	Kuzhora	Kuzhora	*Sindesma*	Kuzhora
Veselyanka	Veselyanka	Veselyanka	Veselyanka	Veselyanka
Sartagan	Sartagan	Sartagan	Sartagan	Sartagan
Ervilia–Pholadidae	*Ervilia*–Pholadidae	*Ervilia*–Pholadidae	*Ervilia*–Pholadidae	*Ervilia*–Pholadidae
Karagan	Karagan	Karagan	Karagan	Karagan
Bryk		Bryk	Bryk	
Yurakov		Yurakov	Yurakov	Toma-kovka beds
Tarkhan *s. str.*		Tarkhan *s. str.*	Tarkhan *s. str.*	
?		Kamyshlak	Kamyshlak	Mayachka Formation
		Korolevo	Korolevo	Karga
		Arabatskaya	Arabatskaya	Chernobaevka
		upper Kerleut	upper Kerleut	Gornostaevka

Regional Konkian Stage. We consider this stage as corresponding in time range to the *Ervilia*–Pholadidae, Sartagan, and Veselyanka beds. There was a long discussion about the stratigraphic position of *Ervilia*–Pholadidae Beds (*Atlas...*, 1959; Molyavko, 1960; Didkovskii, 1964; Barg, 1969, 1993; Koyumdzhieva and Popov, 1985; *Stratigrafiya SSSR...*, 1986), but our recent results substantiated their affiliation with the regional Konkian Stage (Barg and Ivanova, 1998).

The shallow-water carbonate facies of the *Ervilia*–Pholadidae Beds were studied in the central part of the Simferopol' Uplift, where Borehole 1-sk near the Village of Solnechnoe recovered the following beds underlying the lower Sarmatian deposits (below the depth level 66 m):

(1) light gray massive limestone bearing molluscan species *Barnea pseudoustjurtensis* Bog., foraminifers

Miliolinella sp., *Protelphidium subgranosum* (Egger), *Elphidium crispum* (L.), *Cassidulina (?) bogdanowiczi* Konen., *Cibicides* sp., *Globigerina* ex gr. *bulloides* Orb., and accumulations of *Spirorbis* forms (6.5 m thick);

(2) brownish gray solid limestone with mollusks *Ervilia trigonula* Sok., *Spiratella* cf. *andrussovi* (Zhizh.), and *S.* cf. *koeneni* Kittl.; coexisting foraminifers of stratigraphic importance are planktonic *Globigerina bulloides* Orb., *G. praebulloides* Blow, *G. tarchanensis* Subb. Et Chutz., *G. dubia* Egger, *G.* cf. *falconensis*, and benthic *Eponides probatus* Krash., *Glabratella imperatoria* (Orb.), and *Neobulimina elongata* (Orb.); associated fossils are bryozoans and echinoids (0.5 m thick).

The beds overlie the brown-gray Karaganian limestone with *Spaniodontella pulchella* Baily.

Borehole 2-sk (the site 2 km to the south of the village of Novoselovka) penetrated the bed of light gray, organogenic-detrital limestone (2.1 m thick), where *Spirorbis* forms and mollusks *Barnea pseudoustjurtensis* Bog. coexist with rare foraminifers *Quinqueloculina* sp., *Triloculina* sp., *Guttulina* sp., *Fissurina* sp., and *Globigerina* ex gr. *bulloides* Orb.

In the northeastern slope of the Simferopol' Uplift (Borehole 42 located 3 km westward of the village of Zernovoe), the *Ervilia*–Pholadidae Beds of dark gray silty clay, characterizing the relatively deep sedimentation setting, underlie the Sartagan Beds and are 14 m thick. The beds yielded abundant foraminifer species *Discorbis kartvelicus* Satanov and less frequent *D. effusus* Krash., *D. ukrainicus* Satanov, *D. risillus* Bogd., *Cassidulina (?) bogdanowiczi* Konen., *Miliolinella* sp., and *Spiratella* forms. Foraminifers coexist with abundant radiolarians of *Cenosphaera, Liosphaera, Heliomma (Thecosphaera), Cenellipsis, Ellipsohiphus, Cannartus, Phacodiscus, Druppula*, and *Callocecletus* genera (determinations of P.Yu. Lozynyak). In addition, the fossil assemblage includes diatoms and sponge spicules. The abundant and diverse radiolarians detected for the first time in *Ervilia*–Pholadidae Beds of the Crimean Plain (Barg and Lozynyak, 1997; Barg and Ivanova, 1998) show that the host sediments are correlative with the lower Verbovets Beds of the Kosov Formation in the Ciscarpathian region and also with the Solotvino Formation in Transcarpathian areas (*Stratigrafiya SSSR...*, 1986) and coeval deposits in Poland and Romania (Barwicz-Piskoz, 1969; Dumitrica, 1965).

The most complete section of Konkian Stage is that recovered by Borehole 513 in the Akmanaiskii Isthmus (the site of the village of Yachmennoe). We distinguished here all three stage subdivisions described below (from the top downward).

Veselyanka Beds: light gray silty clay with abundant sponge spicules and diatoms. Indicative fossils are *Acanthocardia andrussovi* (Sok.), *Quinqueloculina consobrina sarmatica* Gerke, *Articulina tenella* Eichw., *Ammonia beccarii* (L.), *Nonion bogdanowiczi* Volosh., and *Elphidium joukovi* Serova. The beds are 35.5 m thick.

Sartagan Beds: alternating layers of clay, silt, and shale. These rocks yielded *Spiratella konkensis* Zhizh., *Miliolinella* sp., *Dentalina spinosa* Orb., *Discorbis kartvelicus* Krash., *Cassidulina (?) bogdanowiczi* Konen., *Globigerina bulloides* Orb., *G.* cf. *tarchanensis* Subb. et Chutz., *Subbotina* cf. *cognata* (Pishv.), *Turborotalia mayeri* (Cushm. et Ell.), *Bulimina elongata subulata* Cushm. et Park., *Bolivina dilatata* Reuss, scarce pyritized radiolarians, echinoid spines, and sponge spicules. The beds are 50.5 m thick.

Ervilia–Pholadidae Beds: dark gray to black clay with mollusks *Barnea pseudoustjurtensis* Bog., *Abra* sp., *Spiratella* sp. and others associated with rare foraminifers (*Nodosaria* sp., *Globigerina* sp., and others), radiolarians, diatoms, and echinoid spines. The beds are 31.5 m thick.

Between the *Ervilia*–Pholadidae and underlying Karaganian strata, there is an interlayer of gravelite bearing the mixed Konkian–Karaganian assemblages of mollusks and foraminifers: *Spiratella* sp., *Abra* sp., *Paphia* sp., *Corbula* sp., *Spaniodontella* sp., *Globigerina* cf. *Bollii* Cita et Premoli Silva, *Discorbis kartvelicus* Krash., *Ammonia beccarii* (L.), and *Elphidium macellum* (F. et M.). The interlayer marks a stratigraphic hiatus, and affiliation of the *Ervilia*–Pholadidae Beds with the Konkian Stage appears to be quite reasonable.

The distinguished shallow- to deep-water facies of the Sartagan Beds are of special interest. Facies of the first type have been formerly known in the Tarkhankut Cape only (*Atlas...*, 1959). Sediments of the second type are established in borehole sections 87, 42, and 50 recovered at the Zolotoe Pole, Zernovoe, and Novozhilovka sites, respectively (Barg and Nosovskii, 1993).

In the central Simferopol' Uplift (Borehole 5-sk located 4 km to the northeast of the village of Chaikino), a bed of recrystallized *Nubecularia* limestone (depth interval 47.0–47.6 m) yielded the following fossils: *Nubecularia novorossica* f. *crustaformis* Bogd., *Pyrgo* cf. *affinis* (Orb.), *P. inornata* (Orb.), *Spirolina* sp., *Borelis* sp., *Elphidium aculeatum* (Orb.), *Globigerina* sp., large *Spiratella* and *Spirorbis* forms, solitary corals, and echinoid spines. A similar assemblage is characteristic of the Sartagan Beds recovered by Borehole 2-sk near the village of Novoselovka, where predominant species are *Borelis melo* (F. et M.) and *B.* cf. *pilus* Serova. Coeval sediments of the Al'ma depression (Borehole 42, the village of Nikolaevka) bear the abundant and diverse miliolids (representatives of genera *Quinqueloculina, Triloculina, Spirolina*, and *Hauerina*).

In the Akmanaiskii Isthmus (Borehole 513, the village of Yachmennoe), the studied sequence of Sartagan deposits is the thickest one.

The Veselyanka Beds are detected at many sites of the Crimean Plain region. They bear an impoverished fauna typical of brackish-water environments and indicative of salinity disturbance in the basin of sedimentation. In the Akmanaiskii Isthmus (Borehole 513, the village of Yachmennoe), these are the light gray clay beds occupying the depth interval of 115.5–151.0 m and having a minor silt admixture. In the Al'ma depression (Borehole 6-sk, the village of Prudovoe), the beds are composed of light gray organogenic limestone occurring at the depth from 126 to 139 m. In the eastern slope of the Simferopol' Uplift (the village of Naidenovka), deposits of the Veselyankan age correspond to the relatively thin (2 m) bed of light gray clay. Near the Gulf of Sivash (the Krasnoperekopsk area), the beds under consideration are 6.6–19.1 m thick and show intercalation of recrystallized limestone and sandstone. Pishchikova (1992) distinguished coeval beds of organogenic-detrital limestone (16 m thick in total) near the Dzhankoi.

In general, the Veselyanka fauna from the study region includes the following species of mollusks and foraminifers: *Plicatiforma praeplicata* (Hilb.), *Corbula gibba* Olivi, *Ervilia trigonula* Sok., *Alveinus hitidus* (Reuss), *Acanthocardia andrussovi* (Sok.), *Mutilaster volhynicus* (Eichw.), *Venerupis vitaliana* (Orb.), *Quinqueloculina* ex gr. *consobrina* (Orb.), *Q. consobrina sarmatica* (Gerke), *Q.* ex gr. *reussi* (Bogd.), *Q. reussi sartaganica* Krash., *Articulina tenella* (Eichw.), *Elphidium macellum* (F. et M.), *E. joukovi* Serova, *Ammonia galiciana* (Putrja), *A. pseudobeccarii* (Putrja), *Nonion bogdanowiczi* Volosh., *Bulimina elongata* Orb., and some others.

The listed fossils from Konkian deposits of the Crimean Plain imply that their host rocks are coeval with the upper Badenian sediments known in the western Paratethys (Andreeva-Grigorovich *et al.*, 1995).

Regional Sarmatian Stage. Terrigenous–carbonate and clayey deposits of this regional stage are widespread in the Crimean Plain region. Sediments of the last type characterizing sedimentation in the rather deep settings mostly occur in eastern areas of the region and represent constituent rocks of the thick Krasnoperekopsk Formation corresponding in age to the early–middle Sarmatian. Data on studied foraminifers are used to subdivide the Sarmatian deposits not only into the Volhynian, Bessarabian, and Khersonian substages, but also into units of a lower rank, like beds or horizons in interpretation of Didkovskii (1964).

Lower Sarmatian (Volhynian) Substage, Kuzhora Beds. The beds in question overlie the older deposits of the Konkian Stage either discordantly (the central Simferopol' Uplift), or concordantly (area of the Gulf of Sivash, the Al'ma depression, and Akmanaiskii Isthmus). In the first case, assemblages of euryhaline *Nonion, Elphidium*, and *Protelphidium* species quickly replace the assemblage of *Borelis, Pyrgo*, and *Spiratella* forms across the boundary between the Konkian and Sarmatian strata. In other areas, the late Konkian and early Sarmatian oryctocoenoses are fairly similar (common *Articulina, Elphidium*, and *Ammonia* forms), and the Veselyanka beds are distinguishable here owing to presence of rare *Bulimina* and *Bolivina* species typical of the Konkian level. The Kuzhora Beds vary in thickness from 1–5 m in the Simferopol' Uplift to 33 m in the Akmanaiskii Isthmus.

The overlying *Zbruch Beds* differ in lithology from the Kuzhora beds in the Al'ma depression and central Simferopol' Uplift. In the Akmanaiskii Isthmus, the beds are recognized owing to the almost complete disappearance of *Quinqueloculina karreri* Reuss and its subspecies, and also due to the growing abundance of *Elphidium* forms. The beds are 5.0–25.0 m thick.

Middle Sarmatian (Bessarabian) Substage. In homogeneous sequences of clayey and carbonate deposits, the boundary between lower and middle Sarmatian substages is placed at the first occurrence level of *Dogielina* cf. *kaptarenko* Bogd. et Didk. (Akmanaiskii Isthmus), *Articulina apscheronica* Bogd., *Protelphidium aragviense* (Djan.) (Simferopol' Uplift), and *Spirolina* cf. *pseudostelligera* Didk. (Gulf of Sivash area).

The Novomoskovsk Beds of the substage yield oryctocoenoses showing coexistence of *Quinqueloculina reussi* Bogd., *Q. sarmatica* Karrer, and abundant elphidiids with the following new taxa: *Dogielina* cf. *kaptarenko* Bogd. et Didk., *Articulina apscheronica* Bogd., *Quinqueloculina arcuata* (Didk. et Gudina), *Q. delicatula* (Kolesnik.), *Q.* ex gr. *voloshinovae* (Bogd.), and *Triloculina* aff. *ukrainica* Serova. The beds are 0.4–20.0 m thick.

The Vasil'evka Beds clearly differ from older strata owing to abundance of foraminiferal genera *Dogielina, Sarmatella, Meandroloculina*, and *Nubecularia* typical of the middle Sarmatian. The rod-shaped miliolids (*Dogielina, Sarmatella*, and some other forms) predominate in the deep-water facies, whereas the sessile *Nubecularia* species are dominant in sediments accumulated in coastal settings. The beds are 5.0–34.0 m thick.

The Dnepropetrovsk Beds are easily distinguishable in many studied sections owing to their lithological features. The impoverished oryctocoenoses of the beds usually include scarce representatives of *Triloculina, Flintina, Protelphidium*, and *Elphidium* genera. The beds are 5.0–30.0 m thick.

Upper Sarmatian (Khersonian) Substage. Deposits of the substage are almost devoid of foraminifers. Single representatives of these microfossils are usually undeterminable at the species level. They occur together with redeposited Cretaceous to Paleogene foraminifers and scarce gyrogonites of charophytes. In the Akmanaiskii Isthmus, where the interval in question is composed of clayey deep-water facies, we discovered an assemblage of foraminifers with dominant *Discorbis kartvelicus* Krash. A comparable monotypic fauna

that consists predominantly of *Discorbis* forms is also known from the upper Sarmatian deposits of Azerbaijan (Pobedina *et al.*, 1956). Deposits of the substage are 5.0–85.0 m thick.

The upward transition from upper Sarmatian to Maeotian deposits is traceable almost everywhere in the study region. The boundary between these stages is marked by appearance of indicative Maeotian foraminifers.

Regional Maeotian Stage. New data of geological survey in the Akmanaiskii Isthmus revealed two subdivisions of this regional stage. In borehole sections of the Crimean Plain region, the Maeotian Stage is represented by shallow-water carbonate deposits. An exception is the southern margin of the Indol–Kuban' depression, where clayey to sandy sediments are most characteristic of this stratigraphic interval. The sequence of Maeotian deposits in the Akmanaiskii Isthmus (Borehole 501 situated 4.3 km to the northwest of the village of Frontovoe) is most complete as compared to others known in southern Ukraine. At this site, the Kimmerian strata overlie beds of sand and clay, the upper portion of which is 11.2 m thick (depth interval of 138.0–149.2 m) and yields an assemblage of foraminifers *Quinqueloculina seminula maeotica* Gerke, *Q. gracilis* Karrer, *Elphidium fedorowi* Bogd., *E. macellum maeoticum* Gerke, *Ammonia pseudobeccarii* (Putrja), and *Bolivina dilatata* Reuss. This interval is attributed to the local Akmanai Substage. Underlying beds of the local Bagerovo Substage are 21.5 m thick and bear planktonic foraminifers *Globigerina bulloides* Orb., *G. dubia* Egger, *G.* cf. *tarchanensis* Subb. et Chutz., *Globorotalia scitula* Brady, *G. brevispira* (Subb.), *Turborotalia mayeri* (Cushm. et Ell.), and *Globigerinoides* cf. *trilobus* (Reuss) in association with abundant benthic forms typical of Maeotian deposits in the Crimea–Caucasus region (Bogdanovich and Ivanova, 1977).

In the section of Borehole 520 (3 km to the north of the village of Vladislavovka), the lower portion (Bagerovo Substage) consists of greenish gray beds of sandy clay. These 15-m-thick beds yielded the molluscan assemblage typical of normal marine environments: *Dosinia maeotica* Andruss., *Mytilaster volhynicus* (Eichw.), *Ervilia minuta* Sinz., *Loripes pseudoniveus* (Andruss.), *Cerastoderma arcella mithridatis* (Andruss.), *Abra tellinoides* (Sinz.), and *Congeria panticapea* Andruss. In the upper beds of greenish gray clay intercalated with sandstone and siltstone interlayers (Akmanai Substage 40 m thick in total), we identified the brackish-water species *Congeria subnovorossica* Andruss. and *C. amygdaloides* Dunk.

Near the Gulf of Sivash, the Maeotian sequence (40 m thick) of yellowish to light gray beds of organogenic-detrital limestone was recovered at the Vishnevka site (Borehole 107). We divided this sequence into the lower marine and upper brackish-water parts on the basis of distribution of abundant mollusks and foraminifers. Borehole sections 105 (the village of Krasnoarmeiskoe), 104 (the village of Tomashevka), and 102 (the village of Tselinnoe) are of the same type.

The Maeotian deposits in the studied sections are correlative, according to composition of fossil assemblages, with coeval sediments known from other localities of the Crimea–Caucasus region (*Stratigrafiya SSSR...*, 1986). Bogdanovich and Ivanova (1977) identified nannoplankton of the *Discoaster hamatus* Zone NN 9 and *Discoaster neohamatus* Zone NN 10 at the levels of Bagerovo and Akmanai substages, respectively. Using these data, they correlated the Maeotian with the Tortonian Stage of the Mediterranean region and confirmed some earlier conclusions of Il'ina (1980) and Nevesskaya *et al.* (1984).

Using all the results of biostratigraphic investigations, we present below the geological history of the Crimean Plain region during the Miocene.

At the beginning of the late Kerleut (Gornostaevka) time, a marine basin that originated in the early Kerleut (Askanian) period subsequently spread over the study region in response to intense orogenic movements in the northwestern Caucasus. These movements ceased at the end of the late Kerleut time, and the relevant sea regression transformed the Crimean Plain and northern Black Sea regions into the land.

The new short-term transgression phase was triggered in the northwestern Caucasus at the Arabatskaya (Chernobaevka) time, when the Ol'ginskaya basin in northern Ciscaucasia was directly connected with the "Arabatskaya Sea" in Crimea and the "Chernobaevka Sea" in northern Black Sea region. The next regression at the end of the Arabatskaya time resulted in accumulation of the Korolevo Formation in Crimea, Ritsa Formation in Ciscaucasia, and Karga Beds in the northern Black Sea region.

One of important events at the terminal early Miocene was the Tarkhanian (Tomakovka) transgression initiated in the Ciscarpathian region during development of the Carpathian meganticlinorium at the Nagoryany (Balichi) time, when a foredeep that emerged along the outer edge of the Carpathian Mountains was a source of transgression advanced through the Ciscarpathian region into southern Moldova, northern Black Sea region, and Crimea. Sedimentary sequences, which accumulated during this event, correspond to the Nagoryany and Berezhany beds in Ciscarpathian sections, Podolia Formation in Moldova, Tomakovka Beds and Mayachka Formation in northern Black Sea region, and to deposits of the Tarkhanian Stage in Crimea.

That transgression advanced along two seaways. One was situated between Dobruja and Crimean Peninsula; it had a gulf entering the Al'ma depression, where we discovered abundant remains of planktonic foraminifers and nannoplankton. Another branch of transgression followed the deepest localities of the Karkinit depression leaving the Tarkhankut Rise and central

Simferopol' Uplift as land areas. The most complete Tarkhanian sections remained intact in the Kerch Peninsula and Akmanaiskii Isthmus.

During the Yurakov regression phase, the sea retreated from Crimea, northern Black Sea region, and Moldova. The event obliterated connections with the Ciscarpathian region, as none of the sites in the study region exhibits beds similar in full measure to the lower Badenian strata and bearing characteristic planktonic foraminifers *Orbulina suturalis* Bronn. (Barg *et al.*, 1992).

One more transgression in the Crimean Plain region took place during the second half of the Tshokrakian period in response to ascending movements in the meganticlinorium of northwestern Caucasus. As shown (Barg, 1993), there was a considerable continental epoch in the northern Black Sea region, Crimean Plain, and Kerch Peninsula before this event. The same is evident from general tectonic environments in the central Ciscaucasia at that time (Kosarev, 1971). S.Ya. Ramskoi (personal communication) suggests that continental deposits of the Tshokrakian time could be found in the Akmanaiskii Isthmus (between the Vladislavovka and Novopokrovka sites) and in the Kerch Peninsula near the village of Krasnogorka, i.e., between the Korolevo anticline and Parpach cornice.

As we believe, a shallow-water basin of the late Tshokrakian time was much greater in area than the Black Sea at present. This seems to be obvious, because the upper Tshokrakian strata were detected at the sea floor (within the Golitsyn Rise nor far away from Sevastopol') and also in a series of boreholes drilled along the northern margin of the Karkinit depression and in the outer chain of Crimean Mountains.

Afterward, there was the Karaganian transgression into the northwestern Ciscaucasia. This long-term event triggered erosion, and Karaganian deposits overlie older rocks with a considerable stratigraphic hiatus. In the northern Black Sea region, the transgression became perceptible in the second half of the Karaganian period (Nosovskii, 1960), since we know beds of sedimentary dolomites deposited at that time in deepest areas (Kherson and Nikolaev localities) of the Karkinit depression (Barg, 1993). Deposits of this kind mark, as is known, the terminal stages of transgressive cycles (Chekunov *et al.*, 1976).

After the Karaganian Sea withdrawal, the Indol–Kuban regressive basin accumulated the *Ervilia*–Pholadidae Beds. Some researchers consider these beds as terminal in the Karaganian sedimentation cycle admitting simultaneously that the basin was connected at that time with sea basins in central and western Europe, where the water salinity was normal (Savron' *et al.*, 1990; Konenkova and Ol'shtynskaya, 1996).

When tectonic movements in the Carpathian meganticlinorium became active again, waters of the late Badenian basin went through Moldova and southward of Dobruja in the northern Odessa and Black Sea regions, and then into the Crimean Plain areas, where they entered the basin that accumulated the *Ervilia*–Pholadidae Beds and increased salinity there. The transgression peak was apparently at the terminal stage of deposition of the *Ervilia*–Pholadidae Beds, when the basin extended to the southern edge of the Indol–Kuban' depression, where we detected planktonic foraminifers and radiolarians. The *Ervilia*–Pholadidae Beds should be attributed to the regional Konkian Stage as it is evident from close relations between fossils from these, Sartagan, and Veselyanka beds. The Sartagan Beds accumulated in the northern Black Sea region and Crimea under conditions of growing transgression, whereas the Veselyanka Beds represent deposits of the regressing marine basin.

At the terminal middle Miocene, the early Sarmatian sea basin occupied the Crimean Plain region as a consequence of orogenic movements in the Crimean meganticlinorium. The deepest zone of the basin was at that time in the Indol depression, where clayey sediments accumulated.

The middle Sarmatian time marks the transgression peak, when seawaters reached the southeastern part of the Ukrainian Shield in the north and Moldova in the west.

At the early and initial middle Sarmatian time, ecological environments in some areas of Crimea and northern Black Sea region were identical and specific, favorable for accumulation of black bituminous clay of the undivided Krasnoperekopsk Formation. The middle Sarmatian episode of salinity increase in the basin is thought to be a consequence of more arid climatic conditions at that time (Belokrys, 1967). Retreating afterward, the basin gave way to land in some areas of the Crimean Plain. For instance, the Kakhovka Beds of terrestrial origin are known near the eponymous settlement (Molyavko, 1960).

The late Sarmatian time was marked by deposition of dolomite beds in the Crimean Plain. These beds are especially characteristic of the Gulf of Sivash area.

The semiclosed marine basin of the early Maeotian (Bagerovo) time had a decreased salinity and was related in origin presumably with the East Mediterranean Sea (Il'ina, 1980; Bogdanovich and Ivanova, 1997). Sediments of the regional Akmanai Substage accumulated in a regressing sea basin that occasionally experienced influx of normal marine waters. These episodes are recorded by presence of rare stenohaline foraminifers (genera *Heterolepa, Globorotalia*, and *Reussella*).

As for the Pontian sediments of the Crimean Plain, their affinity with the Miocene Series is substantiated only by nannoplankton (Semenenko and Lyul'eva, 1978). The rocks in question are almost barren of foraminifers and yield molluscan assemblages incompatible with those of the Maeotian time (Barg, 1993).

Thus, we would like to stress in the above discussion that the early–middle Miocene sedimentary basins

of the Crimean Plain had western connections mainly with the eastern Paratethys. Taking into consideration some additional data (Goncharova, 1989; Goncharova and Shcherba, 1997; Goncharova and Il'ina, 1997; Nevesskaya *et al.*, 1984, 1986), one should expect as well the possible influx of normal seawaters from the southeastern Paratethys.

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On Connections between Basins of the Eastern Paratethys and Adjacent Seas in the Middle and Late Miocene

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Abstract—The history of hydrologic and faunal changes in the Eastern Paratethys during the middle and late Miocene is briefly outlined. An effort is made to establish the location of straits, through which marine waters penetrated from the adjacent seas with normal salinity. The analyzed distribution of the most diverse polyhaline assemblages of the Eastern Paratethys at different stages of its development showed that the adjacent seas communicated with southeastern areas of the Eastern Paratethys in the early Tshokrakian, middle Karaganian (Varnian time), Konkian (Sartaganian and Veselyanka times), Maeotian, and most likely in the Tarkhanian. Communications with the Western Paratethys took place in the Tarkhanian, Sarmatian, and probably at the end of the Veselyanka period.

Key words: Eastern Paratethys, Tarkhanian, Tshokrakian, Karaganian, Konkian, Sarmatian, Maeotian, mollusks, prochoresis.

The alternation of different basins in the Eastern Paratethys, namely of those with normal salinity to considerably freshened or brackish-water ones, resulted mainly from the changeable character of their connection with the global oceanic waters. Changes of hydrologic regime in the seas affected the composition of their fauna. The easier were communications of the Paratethyan basins with the World Ocean, the more diverse polyhaline fauna inhabited them. The restricted communications with open seas led to the reduced salinity, to the extinction of polyhaline forms and flourishing of marine euryhaline species, and to the appearance of endemics, whose abundance was the greater, the longer was the period of the abnormal salinity regime in the basins. Events of almost complete basin closure resulted in the extinction of marine taxa and in their replacement by brackish-water species capable of inhabiting waters with another ionic composition like in the Caspian-type basins with prevailing ions SO_4^{2-}, Ca^{2+}, etc.

Judging from the latest publications on the middle Miocene fauna of the Eastern Paratethys, the majority of researchers agrees that the most free communications with the World Ocean were characteristic of the Tarkhanian and Konkian stages. The early Tarkhanian sea was actually inhabited by diverse marine mollusks, pteropods included, and also by benthic and planktonic foraminifers, diverse nannoplankton, sea urchins, brachiopods, and corals. The salinity at that time was nearly normal (over 32‰). However, the fauna became slightly impoverished with time due to the restricting connections of the Tarkhanian sea with the open waters, though the basin salinity presumably was not below 30‰ (Goncharova, 1989).

The succeeding Tshokrakian basin was greater to some extent. Despite a rather free communication with the open seas in the early Tshokrakian time, the basin salinity somewhat departed from the normal, but never dropped below 25‰. Abundance of polyhaline molluscan species and nannoplankton diversity substantially decreased at that time. In the second half of the Tshokrakian, the closure and freshening of the sea began that was followed by extinction of many marine taxa.

The Karaganian basin was even more isolated, and its salinity considerably departed from the normal. Benthic fauna of that time was greatly impoverished and endemic, and nannoplankton disappeared (Muzylev and Golovina, 1987). A short-term communication with the Tethyan waters that resulted in the prochoresis of some marine mollusks and scarce nannoplankton species recommenced only in the mid-Karaganian, namely, in the Varnian time.[1] As a result, the abnormal salinity of the Karaganian sea did not change noticeably, but its oscillations led to extinction of the former species and to appearance and predominance of new taxa, such as *Savanella, Archaschenia,* and others. At the next Kartvelian stage, Pholadidae dominated instead of Lutetiidae.

At the beginning of the Konkian ("Sartagan"), the Eastern Paratethys was well connected with the World Ocean, and its salinity was no less than 30‰. At that

[1] The stratigraphic subdivision of the Karaganian is that accepted in *Stratigrafiya SSSR*, 1986.

time, the basin was inhabited by marine mollusks, many polyhaline species included, and also by bryozoans, various benthic and planktonic foraminifers, ostracodes, and by diverse nannoplankton; echinoderms and corals occurred in the sea as well.

The lifetime of the basin was short. The gradual restoration of communication with the open seas resulted in formation of the semimarine Veselyanka basin inhabited by euryhaline species, many of which were endemic. However, in the second half of the Konkian, communications with the open seas improved again two times at least (Il'ina, in press). In the latest Konkian, a short-term ingression of seawaters occurred in the Transcaspian region (western Turkmenistan), Georgia, Ciscaucasia, Don River lower reaches, and the Sea of Azov region (Il'ina, 1993). The hyperhaline mollusks, namely, *Turriitella*, large *Natica, Murex*, and others invaded the basin together with abundant *Helicosphaera kamptneri* Hay et Mohler known from western Georgia (Minashvili, 1986).

In response to substantially limited connections with the open seas, a new freshening in the basin began after the ingression. A considerable impoverishment of all marine faunal groups, families, and genera occurred in the early Sarmatian. In addition to survived late Badenian and Konkian euryhaline species of mollusks, there appeared many new endemics (Paramonova, 1994; Il'ina, 1998). Only in the west of the basin, its Vienno-Pannonian, Dacic, and, partially, Euxinian parts were still inhabited by the so-called hemistenohaline species (the term proposed by E. Koyumdzhieva, 1969) in the early and at the beginning of the middle Sarmatian (Roshka, 1987). The impoverished assemblages of planktonic foraminifers (Bobrinskaya and Kurenkova, 1986) coexisted with diverse nannoplankton (Muzylev and Golovina, 1987) at that time in the western Paratethys. At the beginning of the middle Sarmatian, the new Mediterranean immigrants appeared among benthic foraminifers as well (Venglinskii, 1975). All this suggests that the Paratethys retained its communication, though rather limited, with the open ocean during the early and, partly, in the middle Sarmatian. If these connections were missing, the hydrology of the sea, at least its salinity, would have changed, the marine components would have been extinct, and the brackish-water forms, namely, Dreissenidae among Bivalvia and Neritidae and Pyrgulidae among Gastropoda, would have been predominant. In other words, the Sarmatian sea would closely resemble the Pannonian lake-like sea after its closure in the middle Sarmatian, or some other brackish-water basins of the Caspian type. The highly peculiar and endemic character of the early and middle Sarmatian fauna (Paramonova, 1994; Il'ina, 1998) resulted most likely not from the complete isolation of the basin, but from its development during a considerable time span (about 2 m.y.) under conditions of a weak water interchange with the World Ocean. The monotonous and substantially poor composition of mollusks and other faunal groups suggests, together with the absence of nannoplankton in the late Sarmatian sea, the partial or complete disruption of communications between the Paratethys and the World Ocean.

Connections of this kind recommenced in the early Maeotian. The communication most likely occurred via a series of intermediate basins, because the early Maeotian sea was inhabited beginning from the onset by a highly endemic fauna (Il'ina *et al.*, 1976). Unmodified components penetrated into the Euxinian–Caspian region (western Georgia) during the transgression maximum in the second half of the early Maeotian (Il'ina, 1980; *Istoriya...*, 1986). At that time, pteropods, including *Spiratella*, coexisted in the basin with a fairly diverse nannoplankton (Semenenko *et al.*, 1995). The diverse assemblage of marine mollusks is also recorded in the Dafni Formation in the Strimon River basin, northeastern Greece. However, many brackish-water forms were encountered there together with some polyhaline and euryhaline forms (Stevanovich and Il'ina, 1982).

In the late Maeotian, the basin was isolated from the open waters. Its fauna was dominated by brackish-water *Congeria, Theodoxus, Turricaspia, Pseudamnicola*, and others. A short-term ingression occurred however at that time. Its traces are recorded in western Georgia and the Kerch Peninsula as rare findings of euryhaline marine forms *Sphaeronassa, Mactra*, and some others (Il'ina *et al.*, 1976; Il'ina, 1980). In the Pontian, a stable brackish-water regime was set up in the Paratethys. Nevertheless, the euryhaline *Parvivenus widhalmi* (Sinz.) penetrated into the early Pontian Euxinian–Caspian basin, suggesting a new connection phase with the true marine waters. This was the general description of hydrologic changes (Fig. 1) and character of faunas in the Eastern Paratethys during the middle and late Miocene.

Although the periodic water exchange between the Eastern Paratethys and the World Ocean is beyond question, the location of straits favorable for this exchange and prochoresis of marine faunas is still unclear and controversial. In order to discuss this problem as applied to the Eastern Paratethys, we should consider first some data on the modern Black Sea.

Having salinity of 17–18‰, the Black Sea is inhabited by impoverished euryhaline marine fauna of the Mediterranean subtropical biogeographic region (Golikov and Starobogatov, 1968). Among the mollusks, about 80 species of bivalves and over 85 species of shelled gastropods are encountered there. However, 20 bivalve species and 16 gastropod forms occur only in the Bosporus region, i.e., close to the influx area of Mediterranean waters, where salinity is the highest (20‰) for the Black Sea. The echinoderms, namely, the starfish *Mathasterias glacialus* (L.,), the sea urchin *Echinocyamus pusillus* (O. Mull.), and marine ophiuroids, were recorded only in the Bosporus region

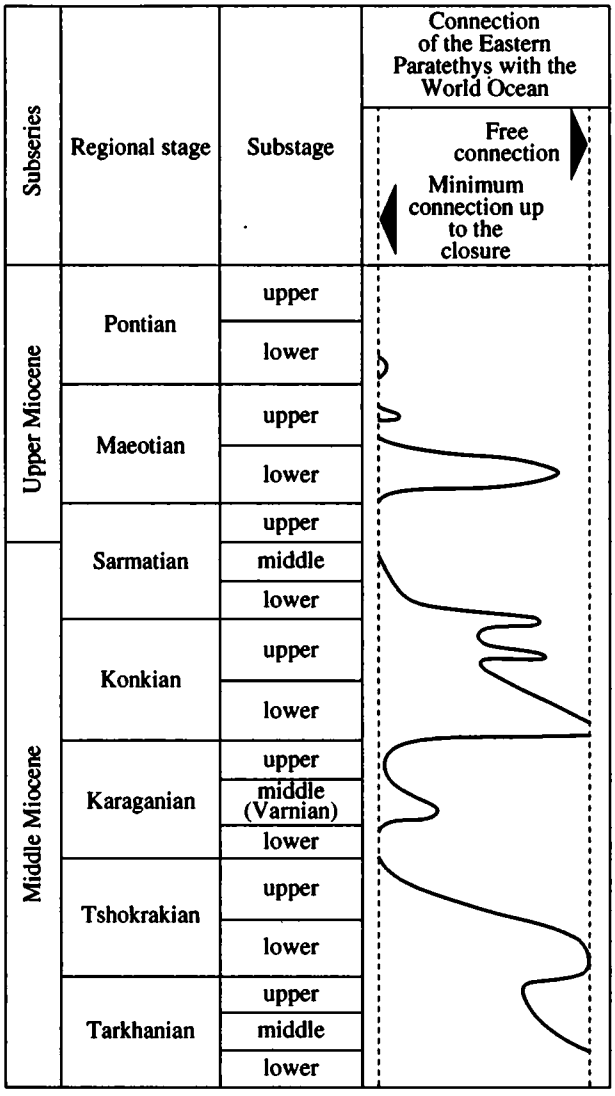

Fig. 1. Changes in communications between the World Ocean and basins of the Eastern Paratethys during the middle and late Miocene.

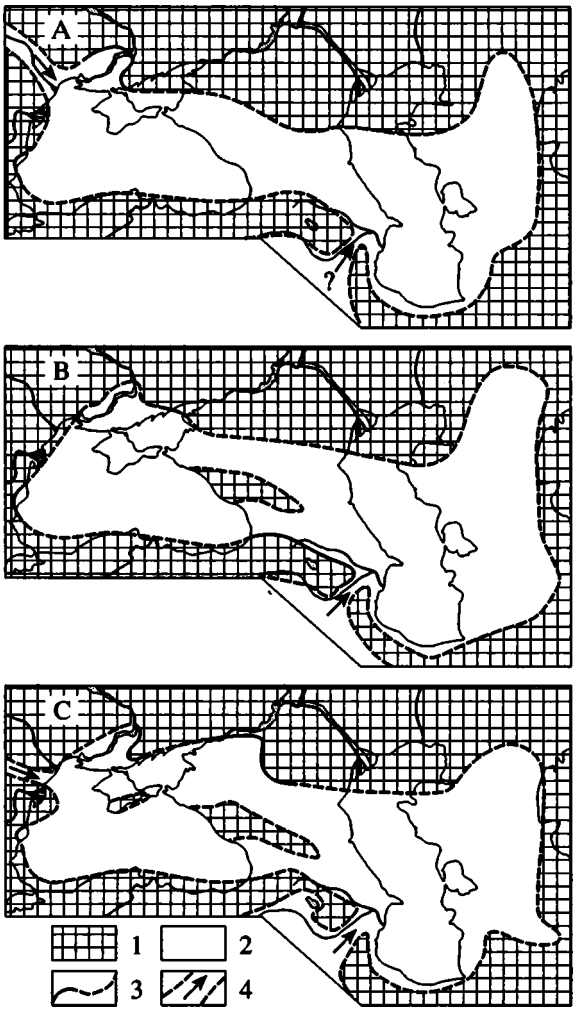

Fig. 2. Presumable straits joining the Eastern Paratethys with the adjacent seas: (A) the Tarkhanian time (after Goncharova and Shcherba, 1997); (B) the Tshokrakian time (after Goncharova and Shcherba, 1997); (C) the Konkian time; (1) land; (2) sea; (3) shoreline; (4) strait.

(*Opredelitel' fauny*, 1972). These animals are absent in other regions of the Black Sea.

It is obvious that polyhaline species of epicontinental seas with abnormal salinity are closely associated with the regions, where straits provide supply of saline water and enable prochoresis of marine taxa. Consequently, the straits that provided penetration of marine waters into seas of the Eastern Paratethys may be detected on the basis of distribution of most diverse polyhaline faunas.

Let us begin with the Tarkhanian period. The Tarkhanian sea was most likely connected with the open waters in the north-west (Fig. 2, A). The past strait joining it with the Western Paratethys may be suggested in the area of the northern Black Sea region (Goncharova, 1989). In the Dnieper River region, namely, nearby the Tomakovka village, 40 bivalve and 20 gastropod species were encountered. The mollusks are represented here by many polyhaline gastropods, such as *Turritella, Nassarius, Aporrhais*, and *Clanculus* that are taxa in common for the Carpathian and Badenian assemblages. At the same time, Goncharova assumed that some polyhaline molluscan species known from the Tarkhanian deposits of the Ustyurt, Georgia, and western Ciscaucasia migrated through the Zagros[2] and Middle Araks straits (Goncharova and Shcherba, 1997). The data on other groups, on gastropods in particular, are as yet unknown.

The distribution of Tshokrakian mollusks, mainly of gastropods, suggests that the communication of the early Tshokrakian sea with the open waters could occur

[2] It is the lame name, the "Upper Euphrates" is more correct.

precisely in the southeast (Fig. 2, B). The most diverse gastropod assemblages are known from the Ciscaucasia, Kerch Peninsula, Georgia, and from the Transcaspian areas, including the northern Kara-Bogaz-Gol region, the Krasnovodskii Peninsula, and the Geokoba Ridge in western Kopetdag. The diversity of gastropods sharply decreases westwards. It was in the early Tshokrakian time, when representatives of the Indo-Pacific genus *Obtortio* first appeared in the Eastern Paratethys and became widespread in the western and central Ciscaucasia, Kerch Peninsula, Georgia, Ustyurt, northern Kara-Bogaz-Gol region, Krasnovodskii Peninsula, and western Kopetdag (Il'ina, 1993, 1995). The representatives of the genus are still unknown west of the Kerch Peninsula thus suggesting the absence of connection of the Tshokrakian sea with the open waters in the west. The communication most likely occurred in the southeast again via the Zagros and Middle Araks straits, as was suggested by Goncharova and Shcherba (1997). The Tshokrakian sea was not directly connected with the Indo-Pacific, but the communication likely occurred via the Eastern Mediterranean that contributed a lot of other Mediterranean gastropods to the eastern Tshokrakian sea (Il'ina, 1993).

A new prochoresis occurred in the mid-Karaganian time, when marine mollusks close in taxonomic composition to the strongly impoverished Konkian assemblage reappeared in the eastern Varnian sea (Mangyshlak, Ustyurt, Georgia, and Ciscaucasia). And again their distribution mostly in the east of the basin suggests the strait location somewhere in the southeast, likely in the same area as in the Tshokrakian time.

The diversity of more or less polyhaline mollusks in the so-called Sartagan (earliest Konkian) basin most markedly proves the communication with the open waters in the southeast. For instance, the assemblage of 42 mollusk species from the site near the Agara station, eastern Georgia, includes 30, i.e., over 70% of polyhaline forms. Afterward, in the second half of the Konkian during the new prochoresis of the Mediterranean fauna established in the Tul'skii settlement, Mangyshlak, and in other sites, the content of polyhaline mollusks was lower (from 45 to 52%). In both cases, however, the diversity of mollusks was considerably declined in western parts of the basin, and the polyhaline species were scarce among them. In addition, it is difficult to assign one or another assemblage of polyhaline mollusks from western areas of the basin to the initial or later period of the Konkian Stage.

In the Transcaspian region (Kara-Bogaz-Gol), Georgia (Naspere village), Ciscaucasia (Dubrovaya Mountain), Don River lower reaches (Novocherkassk), and in the Sea of Azov region, the new prochoresis of marine mollusks, including *Turritella* almost unknown before the latest Konkian time, is recorded. Their assemblages include a lesser number of polyhaline forms than those from the basal Konkian deposits. For example, the Ciscaucasian assemblage includes 15 polyhaline species among 46 forms, whereas only 6 taxa of this kind were encountered among 29 species from the Don River lower reaches. These marine assemblages are associated with the eastern and southeastern parts of the basin, and they again suggest its communication with the World Ocean in the southeast (Fig. 2, C). At the same time, the presence of polyhaline microfauna and nannoplankton in the Konkian sediments underlying the lower Sarmatian in western areas of the Eastern Paratethys (Barg and Nosovskii, 1993; Muzylev and Golovina, 1987) may also indicate the connection between the terminal Konkian sea and Western Paratethys via the Dobruja and Byrlad foredeeps during the late Badenian time (Didkovskii and Nosovskii, 1975).

Therefore, the confinement of most diverse polyhaline faunas to eastern parts of the early Tshokrakian, Varnian, and Konkian basins suggests that the Transcaucasian, Transcaspian, and Ciscaucasian areas were under greater influence of waters with normal salinity. Thus these basins of the Eastern Paratethys communicated with the open sea precisely in the southeast.

The likely communication via Iran and Turkey was suggested by Zhizhchenko (1947), Danov (1957), and Sudo (1961) for the Konkian (Sartaganian) time. Paleogeographic schemes reconstructed for that time by mentioned researchers are also based on the considerable diversity of polyhaline mollusks in the Transcaspian areas. We suppose that the dispersal of relatively polyhaline species mostly in southeastern and eastern parts of the middle Miocene basins of the Eastern Paratethys could result also from the major cyclonic (anticlockwise) direction of currents in the epicontinental seas (Knipovich, 1938).

Based on the available data, is difficult to establish the location of straits that joined the Sarmatian sea with the open waters. We can state only that the Western Paratethys was stronger influenced by marine waters than other areas. It was inhabited by polyhaline mollusks *Crassostrea gryphoides* (Schlot.), *Gastrana fragilis* (L.), *Brachidontes marginatus* (Eichw.), *Clavatula doderleini* (M. Hoern.), *Ocinebrina sublavata striata* (Eichw.), *Mitrella scripta* (L.), and others (Roshka, 1987; Paramonova, 1994; Il'ina, 1998). The planktonic foraminiferal assemblages and nannoplankton more diverse than in the Eastern Paratethys also dwelled there.

The marine molluscan assemblage of the early Maeotian sea that included typical polyhaline Mediterranean species *Gibberula philippi* (Monterst.), *Rissoa ventricosa* Desm., and *Alvania montagui* Payr. is known from the Rioni bay localities of western Georgia, where it coexists with fairly diverse nannoplankton. Like in the middle Miocene, the communication between the Maeotian and Mediterranean basins could occur somewhere in the southeast, possibly, across the modern territory of eastern Turkey or Iran (Il'ina, 1980;

Istoriya..., 1986). On the other hand, a considerably diverse assemblage of marine Maeotian mollusks found in the Strimon River basin can suggest connections between the Mediterranean and Paratethys within the area of Greek Macedonia (Stevanovich and Il'ina, 1982). However, if we admit the communication via the Aegean region, it remains unclear why the euryhaline *Parvivenus* forms did not migrate in the Euxinian–Caspian sea during the Maeotian and appeared in the Eastern Paratethys as late as in the early Pontian only. The short-term late Maeotian ingression most likely was not through the western pathway that opened only in the early Pontian. This is evident from the prochoresis of *Parvivenus* forms.

The established time and ways of the marine fauna prochoresis refine correlation and detail paleogeographic reconstructions. However, the complete resolution of issues under discussion needs further complex investigation of animals that inhabited basins of the Paratethys in the middle and late Miocene. The major problem is an imperfect geologic and stratigraphic examination of southern and southeastern margins of the Paratethys (Azerbaijan, Iran, Armenia, and Turkey border areas). We still do not know answer to the question why pronounced changes in faunas have not been recorded in the Mediterranean area during almost 3.5 m.y., i.e., during the Serravalian correlative to the Karaganian, Konkian, and to a large part of the Sarmatian, when a lot of significant events occurred in the Eastern Paratethys, namely the Varnian, early and late Konkian transgressions, as well as a substantial restriction of communications with the open seas during the early Karaganian and Sarmatian.

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Climatostratigraphic Subdivision of the Moskowian Horizon in the Southeastern Part of the Moscow Area

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Abstract—The Moskowian glacial horizon of the middle Pleistocene is distinguished in the center of the Russian Plain, and, despite a large distribution area, it is poorly studied lacking, for example, the universally adopted regional scheme of its stratigraphic subdivisions. The suggested subdivision of the Moskowian glaciation into several stages is a result of a detailed study of several sections in the Mozhaisk–Tarusa glaciation radius. The best defined are the Borovsk, Kaluga, Mozhaisk and Krasnyi Kholm stages divided by warming phases of different rank. The analytical (lithological–mineralogical and spore–pollen) data, in combination with results of the radiothermoluminescent (RTL) dating, show that the time span between the first and the second (Borovsk and Kaluga) stages was the longest and distinct owing to considerable changes in the spore-pollen spectra of the cool to moderately warm type close to those of the interglacial phases. This interval is ranked and named as the Protvino Megainterstadial. Sediments of this interval are widespread and can be correlated with the deposits of the Kostroma interval recognized earlier in the middle Volga region. The obtained data allows a reliable identifying of three large stages in the Moskowian epoch: the early and middle Protvino, and the late Moskowian stages.

Key words: stratigraphy, Moskowian glaciation, lithology, Protvino Megainterstadial, spore–pollen spectra, palynozones.

The Moskowian Glaciohorizon of the middle Pleistocene spans the upper part of the Quaternary section, and its deposits are widespread in the Moscow region. It corresponds to the uppermost morainal complex overlain only by the thin sedimentary cover. Despite the well-exposed rocks easily accessible to investigation, the Moskowian Horizon and Moskowian glaciation (the time interval of 190 or 200 to 300 ka) is as yet poorly studied. The stratigraphic position of the horizon is fairly definite, but its paleogeographic essence and events of that time period are treated differently. The extreme boundary of the Moskowian glaciation is variably drawn, and different number of glacial stages is recognized by researchers, who differently interpret the sections and paleogeographic events that occurred in the second half of the middle Pleistocene in the Moscow region (Moskvitin, 1946; 1967; Zarrina and Krasnov, 1984; Velichko, 1980; Chebotareva, 1982; Shik, 1957; 1981; 1993).

Moskvitin (1936; 1946) was first to distinguish the Moskowian Glaciohorizon as an individual unit. Using geomorphological data, he identified two glaciation events: the first (maximal) Bronnitsy stage and the second Iksha stage. Shik (1957) substantiated the climatic distinction of the Moskowian glaciation on the basis of paleobotanical data, but this point is still under discussion that also concerns the status of the Moskowian glaciation and of the Dnieper–Moskowian (Odintsovo, Shklov) interval considered as interglacial, interstadial, or megainterstadial. Leaving aside the problem of the Moskowian glaciation status, we must only indicate that the issue was discussed in our previous publications (Antonov *et al.*, 1991; *Kompleksnyi analiz...*, 1992), where we arrived at the conclusion that the Moskowian and preceding Dnieper glaciations represent independent events, and corresponding horizons can be attributed to the middle Pleistocene.

This work was motivated by the fact that studies on the glacial horizons and paleogeographic conditions of sedimentation at the time of glacial epochs was based during many years mainly on geomorphological and lithological–mineralogical data, whereas palynological materials extensively used in studies on interglacial stages were disregarded, as a rule, in works dealing with cold epochs. Only in 1965, Pisareva (1965), who studied the Volga region near Kostroma, described palynological spectra of the lacustrine–glacial beds that divide two layers of the Moskowian moraine. According to her data, pollen of the arboreal–fruticose plants (60–75%) prevail in these beds. Along with pollen of arbuscular birch, pine, and spruce, she detected a considerable amount of pollen grains of fruticose birches (up to 20%) and alders (up to 12%). The spore–pollen spectra also showed presence of lime, elm, wild nut, oak and horn beam pollen (not more than 5–10% in total). Pisareva suggested that pollen of broad-leaved

† Deceased.

taxa was redeposited in the lacustrine–glacial beds. The paleogeographic environment of this middle Moskowian, or Kostroma interval is qualified as rather warm, but somewhat cooler than nowadays. In subsequent papers on the middle Pleistocene epoch (Nosov and Skiba, 1975; Shik, 1981; *Moskovskii lednikovyi...*, 1982), about two dozens of other sections of morainal and intermorainal beds of the Moskowian Horizon were described, and palynological characteristics were studied in the Krasnyi Bor (Smolensk region), Yastrebovo (Vereya), and Pekino localities (the northwestern part of the Moscow region). In general, the intramorainal beds are characterized here by the interstadial climatic environment and show predominance of arboreal pollen with successive alternation of birch–pine and spruce–birch species. Small amounts (up to 10% in total) of pollen of broad-leaved trees (lime, oak, elm, maple, hornbeam) were persistently detected in all samples, but this fact was ignored in the authors' interpretations.

The problem of interpretation of the interstadial pollen floras that include elements of the broad-leaved group is topical for the extraglacial zone as well. For instance, data on sediments constituting the structure of the third terrace in the Oka River valley near the village of Krutitsy (Zelikson and Monoszon, 1981), show that they accumulated prior to the maximal stage of Moskowian glaciation. The authors indicated that the interstadial vegetation was represented by forests with the common and Siberian pines associated with birch and larch. However, they stated that the scarce pollen of broad-leaved species in these sediments was redeposited from the upper Pliocene beds.

In the book mentioned above (*Moskovskii lednikovyi...*, 1982), the sections of intermorainal deposits in the Central Russia Plain that were described and treated earlier as sediments of the Dnieper–Moskowian (Odintsovo) interval (Pogulyaev, 1955; *Rel'ef i Stratigrafiya...*, 1961; Chebotareva, 1959; Danilova, 1959; *Razrezy lednikovykh...*, 1977) were revised, and all the layers were attributed to the Moskowian Horizon. Shik (1981) correlated sediments of the Kostroma interval with the so-called "Bezhetsk" interglacial deposits previously distinguished in the Volga River upper reaches. He also suggested the interglacial status of this interval, thus implying the concept of three independent glaciations of the middle Pleistocene time. Later, however, this idea was not justified (Shik and Biryukova, 1989). The interglacial status of these deposits (namely, of the Barmino Interglacial with the isotopic date of 190 ± 20 ka) was later supported, with reference to Shik's data, only by Zubakov (1992).

The regional character of the Kostroma Interstadial, is sufficiently well confirmed, as Grichuk (1989) and Yelovicheva (1985) stated, by the paleobotanical evidence obtained from a series of sections studied in Belorussia and in the central part of the Russian Plain.

In the same period of time, studies on the structure of glacial deposits and on the lithology and mineral composition of heterochronous moraines in the Russian Plain (Gaigalas, 1979; Astapova, 1978; Sudakova, 1990; a.o.) revealed peculiar features in the composition of the moraines, their main provenances, and displacement direction of sediments by the moving ice. After a comprehensive analysis (granulometry, petrography of rock fragments, mineralogy of heavy and clay fractions, paleomagnetic study of morainal silts, and isotopic dating), the concrete criteria for the diagnosis of glacial complexes different in age were obtained. This approach was used for interregional correlations of glacial deposits, age correction of beds in different sections, and more precise evaluation of time spans of particular stratigraphic horizons. The analysis outlined the most representative areas for the study of different glacial rock complexes: the Dnieper River basin, Volga River area near Yaroslavl, and also the southern and southwestern areas of the Moscow region (Fig.1). Actually, the study of the latter (the Mozhaisk, Borovsk, and Tarusa groups of sections) revealed three lithological–mineralogical varieties of the Moskowian moraine (M_0, M_1, and M_2, from the top downward) corresponding to different stages of the Moskowian glaciation (*Razrezy lednikovykh...*, 1977; Anan'eva *et al.*, 1979; Sudakova and Bol'shakov, 1980; Sudakova, 1990). Sediments of the lowest (M2) moraine were attributed to the Kaluga stage of maximal glaciation, whereas the following two stages (M1 and Mo), or the Borovsk and Mozhaisk morains were considered as traces of consecutive contraction of the ice sheet. Afterward, the detailed observations in the Protva River basin provided data correcting to a certain extent this concept (Antonov, 1990; *Kompleksnyi analiz...*, 1992). For instance, the limited distribution of moraine M2 at the tops of the interfluve areas, where, to the contrarily, M1 deposits are widespread, allowed us to connect the latter with the maximal stage of the Moskowian glaciation.[1] Moreover, the rather thick (up to 10–12 m) sediments of different genesis (fluvioglacial, lacustrine, etc.) present between the morainal layers in many sections suggest that the accumulation epochs of the moraines were separated by a time interval that was longer than typical interstadials. A comparison between the glacial horizons in the Mozhaisk–Tarusa glacial radius (Fig. 1) revealed traces of one unknown glacial stage of the Moskowian glaciation at the latitude of Krasnyi Kholm and Knyazh'i Gory settlements (Antonov *et al.*, 1995). This Krasnyi Kholm stage evidently correlates with the Iksha stage distinguished by Moskvitin. Therefore, by the beginning of the 1990s, there was a quite definite opinion that several moraines exist within the Moskowian Horizon being divided by sediments accumulated during interstadial or interglacial epochs. The largest Kostroma (or Iksha, after

[1] It should be noted here that some of us do not share this viewpoint.

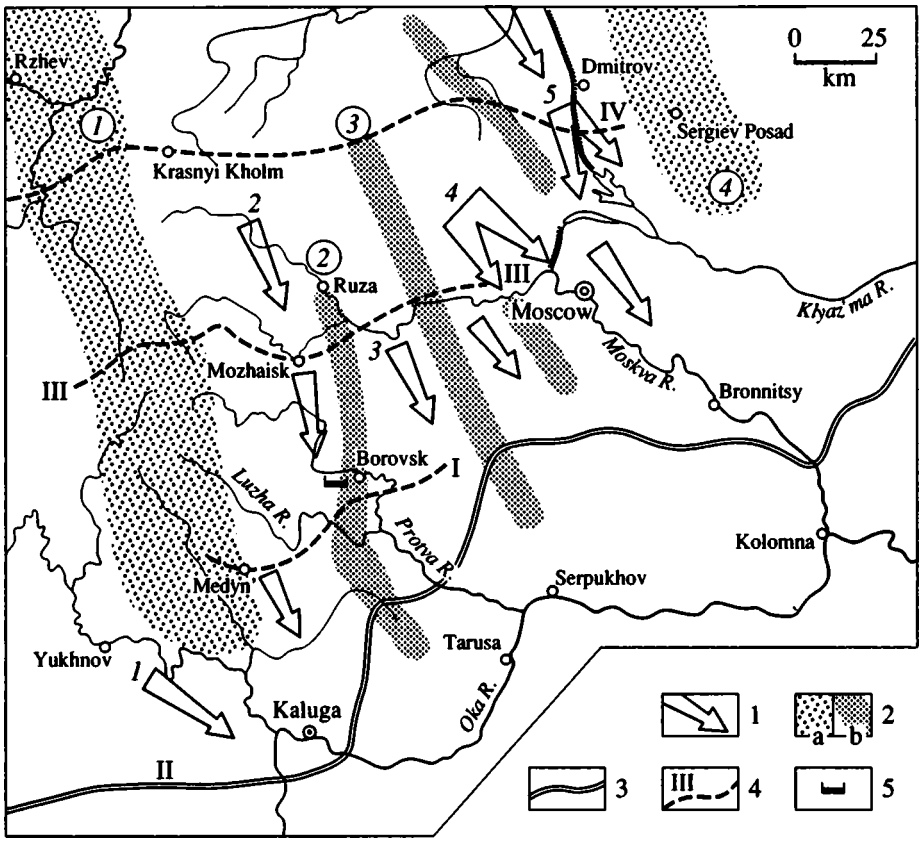


Fig. 1. Structure of the Moskowian glacial sheet in the central part of the Russian plain: (*1*) Ugra, (*2*) Protva–Luzha, (*3*) Nara, (*4*) Istra, and (*5*) Yakhroma glacial flows; (2) ice divide zones of the first (a) and second (b) orders; (3) the maximal boundary of the Moskowian glaciation; (4) the supposed position of stadial boundaries; (5) position of geological profile shown in Fig. 2. Numbers in circles: (*1*) Rzhev–Medyn', (*2*) Ruza–Maloyaroslavets, (*3*) Klin–Voronovo, and (*4*) Sergiev-Posad glacial flows. Stadials: I, Borovsk (moraine M2); II, Kaluga (maximal, moraine M1); III, Mozhaisk (moraine Mo); IV, Krasnyi Kholm.

Nosov and Skiba, 1975) interval was defined as one of the wide regional significance. Because of a considerable lateral variability of glacial deposits, we refrain as yet from correlation of the Moskowian Horizon sections distant from one another and will dwell on data obtained from exposures and boreholes located in a compact group and most favorable for the reliable comparison. The Borovsk group of sections in the Protva River middle courses is of this kind.

In the Protva River basin that is in the central part of the Mozhaisk–Tarusa radius of glaciation, several stadial moraines of the Moskowian Horizon were identified some time ago, and a lot of data on lithology of Quaternary deposits was obtained (Anan'ev *et al.*, 1979; Antonov, 1990; *Kompleksnyi analiz...*, 1992; *Stroenie i istoriya...*, 1996) together with palynological characteristics of sediments accumulated in the upper middle Pleistocene. Until recently, stratigraphic boundaries of glacial horizons in the sections under consideration were mainly determined on the basis of lithological–mineralogical evidences, and sediments of the Moskowian Horizon became characterized in the palynological aspect only now. The obtained results are indicative not only of sedimentation conditions during the Moskowian epoch, but also can serve as independent criteria for drawing the stratigraphic boundaries. In addition, the deposits of the horizon were dated by O.A. Kulikov using the RTL analysis (*Kompleksnyi analiz...*, 1992).

In the Protva River middle courses (Fig. 2), deposits of the Moskowian Horizon are exposed within a wide hypsometric interval ranging from heights in the water divides to slopes of the river valleys. The glacial, fluvioglacial, lacustrine, and alluvial deposits of the horizon overlie the Carboniferous bedrocks and underlying sediments of the Dnieper Horizon being overlain by a thin (1.5–3 m) cover of the middle–late Pleistocene loess-like loam. In the key area of the morainal plain studied in detail, the thickness of the Moskowian Horizon varies from 3–5 to 20–25 m (Fig. 2), whereas, beyond this area and in proximity to terminal moraines, it may reach 35–40 m and more mainly owing to the greater thickness of the morainal deposits (Antonov, 1990; *Kompleksnyi analiz...*, 1992). As a result of our study, we established that the mineralogical spectra of Moskowian moraines almost completely lack the glau-

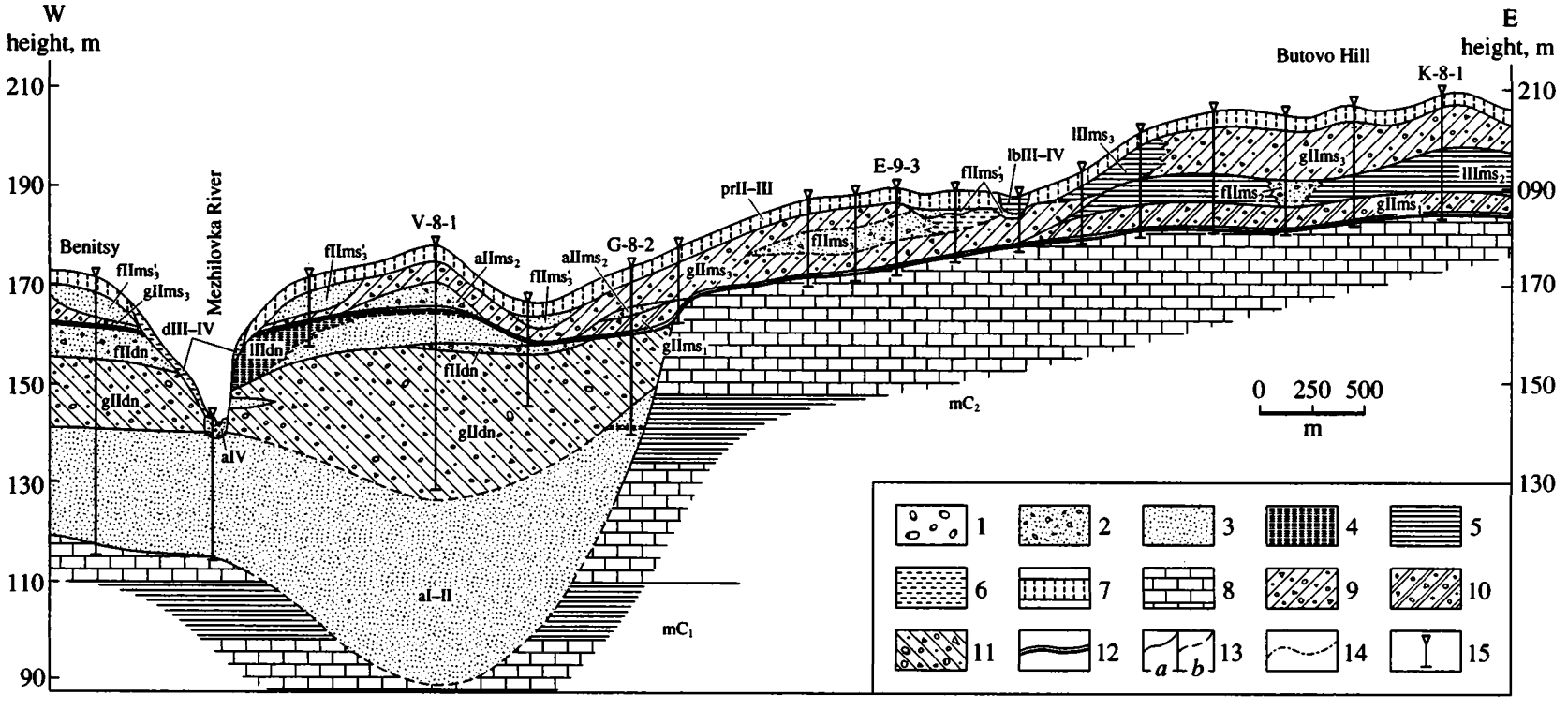
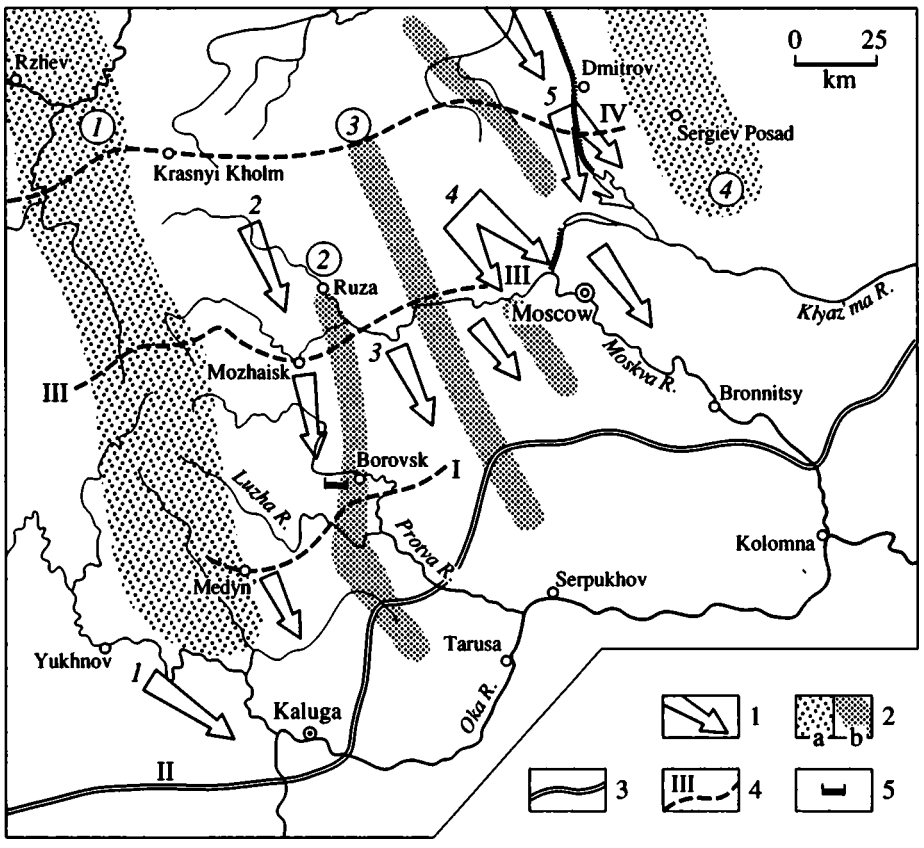


Fig. 2. Geological profile, the right-hand bank of the Protva River, the Borovsk District: (1) boulders; (2) gravel and pebbles; (3) sand; (4) silt; (5) clay; (6) loam; (7) covering loam; (8) limestone; (9) boulder loam (moraine) of the second (Kaluga) stage of the middle Pleistocene Moskowian glaciation (M1); (10) moraine of the early (Borovsk) stage of the Moskowian glaciation (M2); (11) moraine of the middle Pleistocene Dnieper glaciation; (12) lower boundary of the Moskowian Horizon; (13) other stratigraphic boundaries observed (a) and supposed (b); (14) facies boundaries; (15) borehole. Indices for stratigraphic subdivisions: (aIV) Holocene alluvium; (dIII-IV) talus and (lbIII-IV) lacustrine–paludal deposits of the late Pleistocene–Holocene; (prII-III) sedimentary cover of the late–middle Pleistocene; ($fIIms'_3$) upper Moskowian fluvioglacial beds above moraine; ($fIIms_3$, $lIIms_3$, $gIIms_3$) upper Moskowian fluvioglacial, lacustrine, and glacial (M1) deposits; ($aIIms_2$, $fIIms_2$, $lIIms_2$) middle Moskowian alluvial, fluvioglacial, and lacustrine sediments; ($gIIms_1$) lower Moskowian glacial beds(M2); (fIIdn, lIIdn, gIidn) fluvioglacial, lacustrine, and glacial deposits of the Dnieper glaciation; (aI-II) early–middle Pleistocene alluvium; ($mC_{1,2}$) marine sediments of the Early–Middle Carboniferous.



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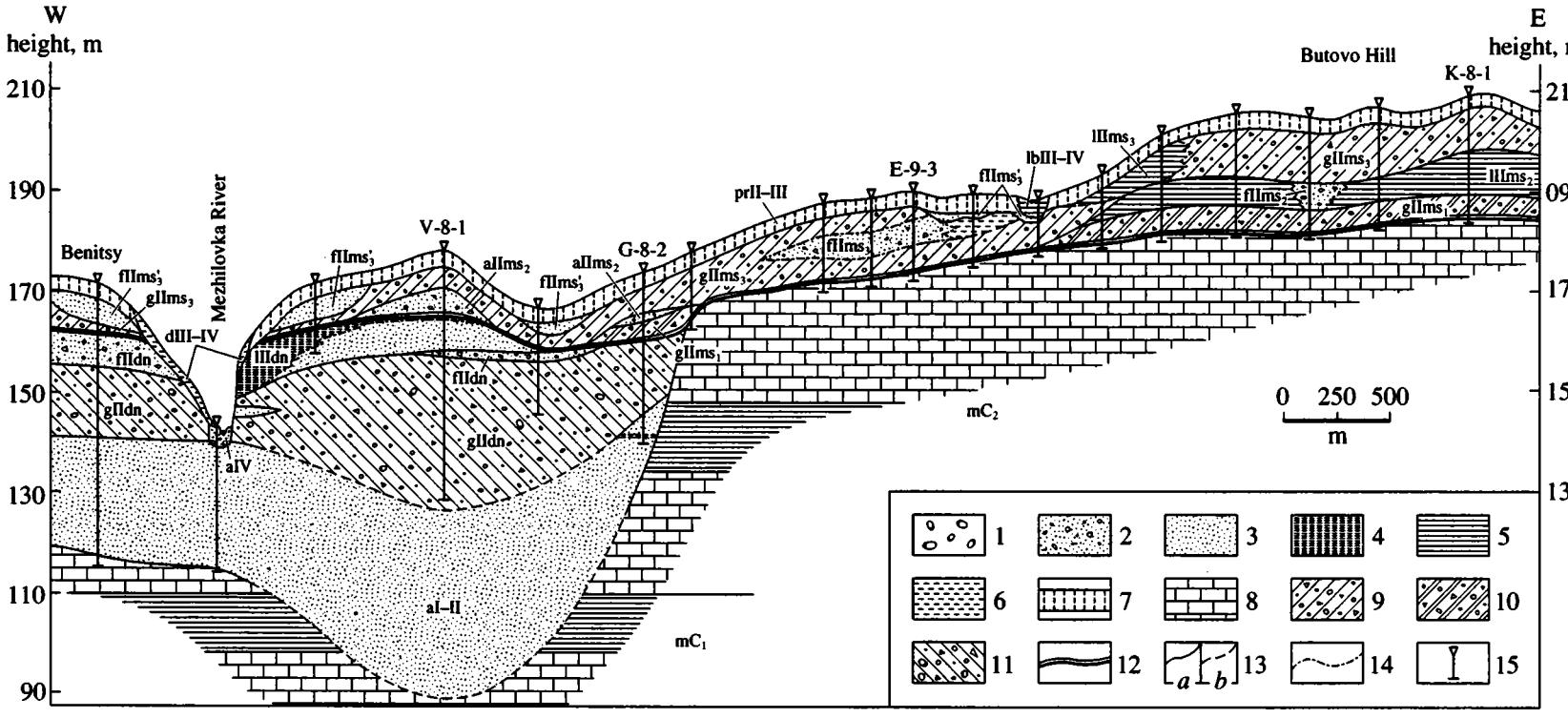

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conite–siderite–sulfide mineral association in contrast to deposits of the Dnieper Horizon. Having the high content of hornblende (up to 25%) and garnet (up to 15%), the Moskowian moraines on the right-hand bank of the Protva River differ in general by their garnet–hornblende mineral association. According to the granulograms, the morainal silts are less clayey here and more sandy than those of the Dnieper level. Mineralogical spectra of fine sandy and coarse silty fractions are fairly rich in components of the Baltic provenance and show the reduced share of local materials (Fig. 3). Analytical data on borehole sections in the right bank of the Protva River between the Benitsy and Butovka villages (Fig. 2) show a considerable spatial variability of glacial deposits and their recurrent wedging and facial replacements along a very limited distance (up to 1 km), thus implying that the permanent lithological–mineralogical control is required when tracing a horizon, or correlating its individual sections. For example, the two-layer morainal section containing the dividing lacustrine and fluvioglacial deposits shows the following lithological–structural peculiarities: sediments of two (Dnieper and Moskowian) glaciations in one case (Borehole V-8-1); three morainal levels (one Dnieper and two Moskowian) in the other (Borehole G-8-2) moreover, the lower Moskowian morain $gIIms_1$ (M2) directly overlies the Dnieper moraine (Fig. 3); contains two layers $gIIms_3$ and $gIIms_1$ (M1 and M2) of the Moskowian moraine in the Borehole K-8-1; and, finally, two beds of the late variety of the Moskowian moraine $gIIms_3$ (M2) in the Borehole E-9-3.

The uppermost (absolute height 209 m) section K-8-1 is located on the apical part of the pre-Quaternary socle elevation and exposes, in addition to two morainal layer, a bed of lacustrine silt and clay accumulated in a dammed periglacial lake. Similar lacustrine deposits were also observed on northern slopes of some elevations in the right bank of the Protva River, westward of the study area. These deposits were likewise found eastward of the area in the vast territory at the left bank of the Moskva River and between the Protva and Nara rivers, where the lacustrine beds rest on the moraine of an earlier stage of the Moskowian glaciation, or between two morainal beds (Isakin, 1976). Thus, the lacustrine beds show a fairly wide distribution, considerable thickness (5–10 m), and distinct stratigraphic position.

The sediments penetrated by the Borehole E-9-3 are confined to a lower interfluve level, and the thin cover of loose deposits above them includes the latest lithomineralogical variety of the Moskowian moraine divided by a lens of fluvioglacial sediments. Finally, boreholes G-8-2 and V-8-1 reached deposits of the Moskowian Horizon in an ancient river valley that was filled in the pre-Moskowian time by thick (20–30 m) beds of sediments, the larger part of which corresponds to the moraine of the preceding Dnieper glaciation and to the pre-Dnieper alluvium. On slopes of the buried valley, the Moskowian Horizon includes both morainal varieties and layers dividing them, whereas in the central part of the valley (Borehole V-8-1), the lower member of the Moskowian moraine was washed out by the fluvioglacial stream and was substituted by beds of pebbles and sand. When tracing these beds over the area, we found that they can be observed along the recent Protva River valley at the altitude of 175–180 m. This situation suggests that the northern slope of the Borovsk–Maloyaroslavets rise near the modern valley of the Protva River was a site of the fluvioglacial runoff. Taking into account the former existence of the dammed lake at a higher hypsometric level of the uplift slopes (see above), we can state with confidence that the sandy fluvioglacial sediments reached by boreholes G-8-2 and V-8-1 are younger than the lacustrine clays of the Butovo Hill. The upper moraine M1 crowning the section includes interbeds and lenses of fluvioglacial sand and loam and shows traces of fluvioglacial reworking during the short-term recessions of the glacier edge.

The uppermost part of the Moskowian Horizon section in the Protva River basin consists of fluvioglacial sandy to pebbly deposits of the runoff dells and outwash plains, and also of loamy lacustrine sediments accumulated in the interfluve depressions. All these sediments are embedded into the moraine M1. According to indirect evidences (geomorphological conditions of bed occurrence, their distribution patterns, and thickness of the intermediate layers), the time spans between the second (maximal) and following stages of the Moskowian glaciation were considerably shorter than the interval separating the first and second stages.

Thus, the lithological, mineralogical, and geomorphological data show that the middle Pleistocene Moskowian Horizon in the basin of the Protva River is divisible into the following climatostratigraphic subdivisions of a lower rank (subhorizons?): the lower Moskowian corresponding to the lower moraine M2; the middle Moskowian coeval with the long-term interstadial; and the late Moskowian correlative with the second glaciation stage that left the moraine M1 and fluvioglacial and lacustrine deposits. To the north of the region, moraines belong to the later Mozhaisk (Mo), Krasnyi Kholm and other glaciation stages.

The detailed palynological analysis of glacial and intermorainal deposits of the Moskowian Horizon revealed important paleogeographic conditions of sedimentation. For example, the lower morainal deposits (ms1, Palynozone 1) recovered by boreholes K-8-1 (Fig. 4) and G-8-2 (Fig. 5), show a rather low content of microfossils and certain predominance of arboreal pollen. They yielded pollen of birch, pine, and alder in association with single grains of the broad-leaved taxa (wild nut and lime). The fruticose group is represented by dwarfed birches and alder, and wormwood, goosefoot, and accessory Gramineae are main herbaceous components.

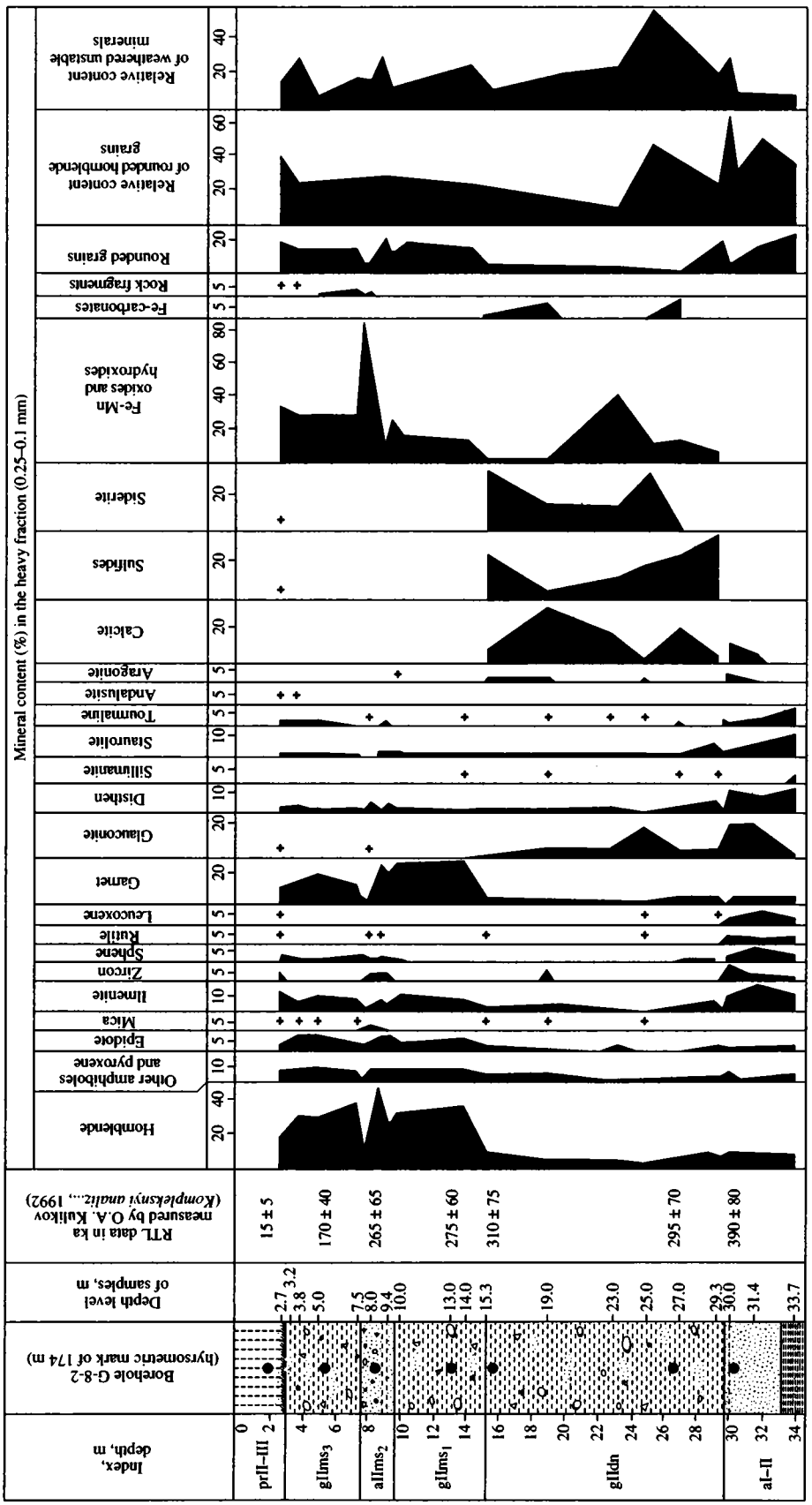

Fig. 3. Mineralogical diagram of deposits recovered by Borehole G-8-2 (symbols as in Fig 2; mineral analysis by E.G. Anan'eva, *Kompleksnyi analiz...*, 1992).

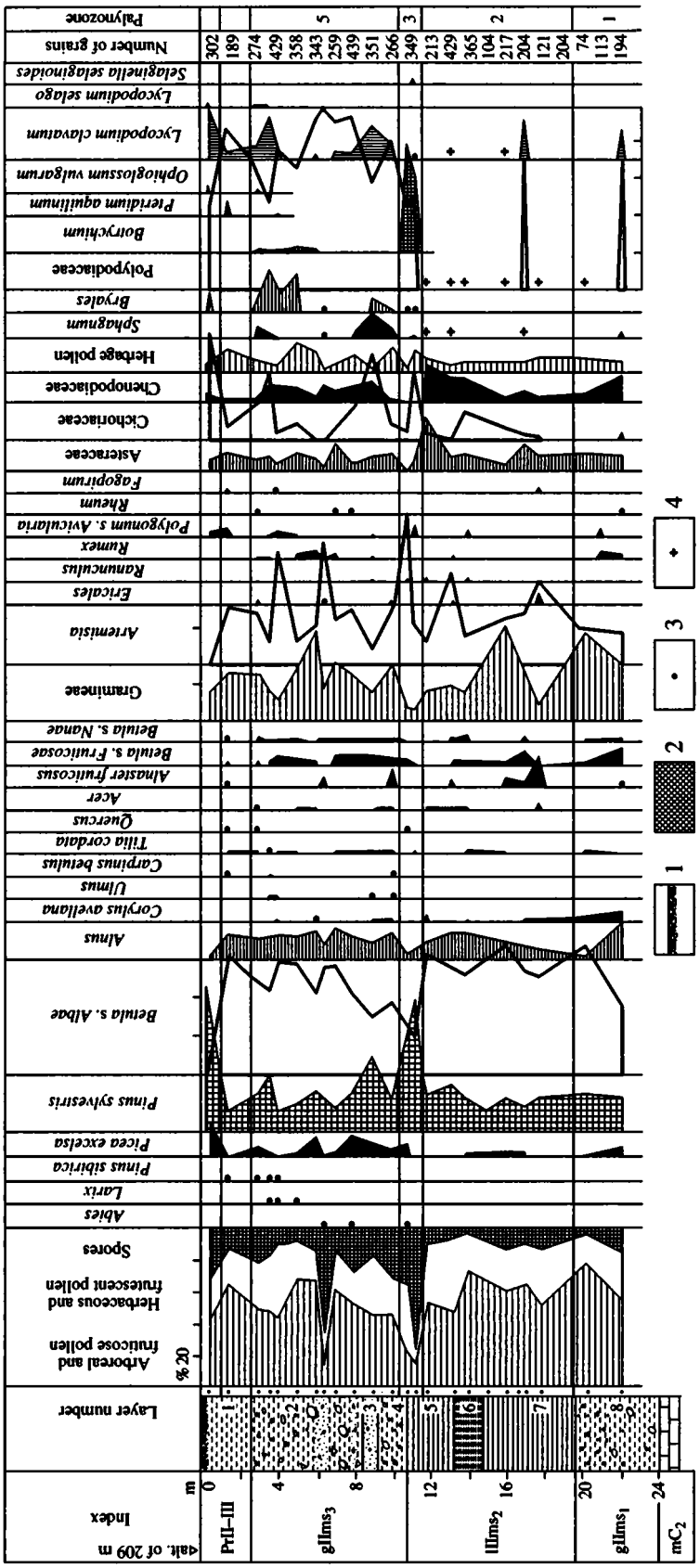

Fig. 4. Spore–pollen diagram for deposits recovered by Borehole K-8-1: (1) recent soil bed; (2) buried humus horizon; (3) pollen-and-spore content less than 1%; (4) single grains (other symbols as in Fig. 2).

The palynological composition of the middle Moskowian intermorainal deposits was studied in three borehole sections (K-8-1, G-8-2, V-8-1), the indicated succession of which elucidates conditions of sedimentation in the Protva River basin during a larger period of the middle Moskowian epoch. The earliest phases of the middle Moskowian Substage correspond to sediments recovered by the Borehole K-8-1 and deposited immediately above the lower morainal level. The spore–pollen data reveal two palynozones in sediments of this interval. The first palynozone (ms2, Palynozone 2) adjacent to the early Moskowian moraine retains the background pollen spectrum of the latter: the pollen of birch, pine and alder still dominates in the arboreal group, though abundance of alder (in some samples), fruticose birch, and heather pollen somewhat grows. In contrast, the next Palynozone 3 shows sudden changes in the pollen and spore composition: the pine pollen share considerably grows, and single grains of oak pollen appear in the spectrum, which also become rich in spores (botryoids, ferns). The indicated changes probably reflect environmental transformations around the sedimentation area in response to draining of the dammed lake and climatic warming.

Higher intermorainal beds reached by the boreholes G-8-2 and V-8-1 (Figs. 5, 6), are, apparently, related to the second half of the middle Moskowian time, and geological data for a part of the epoch under consideration appear to be missing. Palynological data (general composition and separate components) for sandy and pebbly sediments of the horizon that are reached by these boreholes revealed characteristics similar to those typical of the recent alluvium of the Protva River. Samples from layers 5 and 6 of the borehole section G-8-2, like the sample from the layer 5 of the borehole section V-8-1 (Palynozone 4), showed a high concentration of pollen and absence of sudden changes in diagrams of the dendroflora distribution. In these layers, against the background of usual composition of arboreal pollen (pine–birch forest with an admixture of fir and alder), the above layers recorded a slow growth in the content of the broad-leaved species (mainly of lime pollen up to 16–20%). The composition of herbaceous group corresponds to the meadow–forest and meadow assemblages, where Gramineae and forbs are dominant and associate with single pollen grains of aquatic plants. In the upper part of layer 5 (Borehole V-8-1, Fig. 6), an anomalous group, where fern spores represent up to 90%, characterizes the local spore–pollen spectrum that reflects the peculiar local environments of sedimentation in the peripheral part of the ancient valley. In general, the pollen assemblage of Palynozone 4 characterizes the vegetation zone of mixed and broad-leaved forests, and also of the valley forests with alder. In the fossil assemblage, groups of broad-leaved plants and spores are quantitatively more representative than in the present-day alluvium. Obviously, the optimum of the middle Moskowian warming coincided with the widest distribution of the clear birch and pine–birch forests with herbaceous plants and ferns.

The palynological composition of the upper moraine (ms_3, Palynozone 5) considerably differs from that characteristic of the middle Moskowian interval. Coniferous plants, *Pinus sibirica* included, become more abundant among the arboreal pollen, and the content of birch and, particularly, of alder pollen decreases. The broad-leaved taxa are represented by single pollen grains. Concurrently, pollen of fruticose birch and alder appears at this level. The pollen content of herbaceous plants gradually grows from the base of this morainal bed upward (the bed is detected in all four borehole sections); a greater share is characteristic here of sagebrush, composite plants, herbage, and heather. Some samples show larger proportions of spores.

Considering the pollen assemblage of the Moskowian Horizon as a whole, one may note in its composition a peculiar combination of elements typical of mixed forests (birch, pine, spruce, alder, and broad-leaved species) and northern taiga (larch, fir, alder, and fruticose birch).[2] The present-day distribution limit of alder is far away from the study region, and *Alnaster* is characteristic plant in the northeastern part of the European plain and in the northern Urals. In the northern taiga zone, this species is typical component of the underbrush in the coniferous and coniferous-parvifoliate forests, whereas at the northern boundary of the distribution area, it occurs in groupings with *Betula nana*. In the studied sections of the Moskowian Horizon, the alder pollen is found in different combinations with other taxa of dendroflora, but always in the spore–pollen spectra of the forest type. Birch, pine, and alder pollen is dominant in the latter, while the herbaceous group represents the diverse herb–grass meadow and meadow–forest biocoenoses. Fossil floras of this type are typical of the interfluve areas (boreholes E-9-3 and K-8-1). Dominant in these floras is pollen of birch and pine representing "persistent" taxa of the Pleistocene assemblages; permanently present are also representatives of two coenogenetic communities (Zelikson, 1985). One of the latter is the dark coniferous taiga consisting of *Abies, Picea, Pinus sibirica, Licopodium selago, Selaginella selaginoides*, and including the group of nanoid *Alnaster fruticosus, Betula nana*, and *Betula* sect. *Fruticosae*. The second impoverished nemoral assemblage contains the broad-leaved species. The "mixed" fossil floras of this kind were formed, evidently, in the periods of transformation of zonal vegetation. If this is correct, the transitional floras characterize the epochs of cooling and more continental climate.

Fossil floras from the sandy beds in borehole sections V-8-1 and G-8-2 (Palynozone 4) are of the interstadial type proper. However, these beds are fairly thick and widespread (see above), and this suggests a higher

[2] According to Shik, pollen assemblages of this kind may reflect presence of redeposited pollen, and our speculations about paleogeographic conditions at the deglaciation time are debatable.

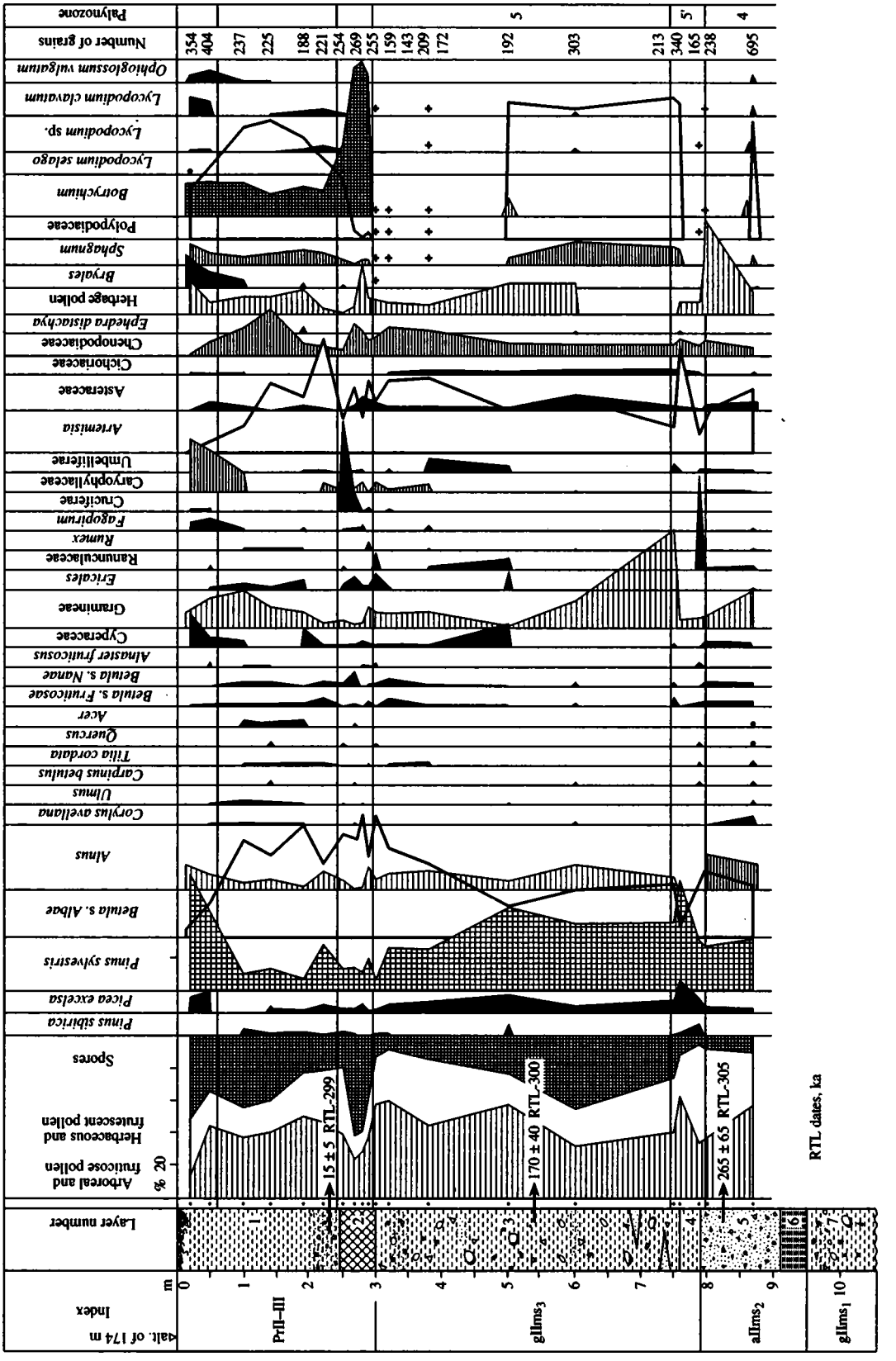

Fig. 5. Spore–pollen diagram for deposits recovered by Borehole G-8-2 (symbols as in Figs. 2 and 4).

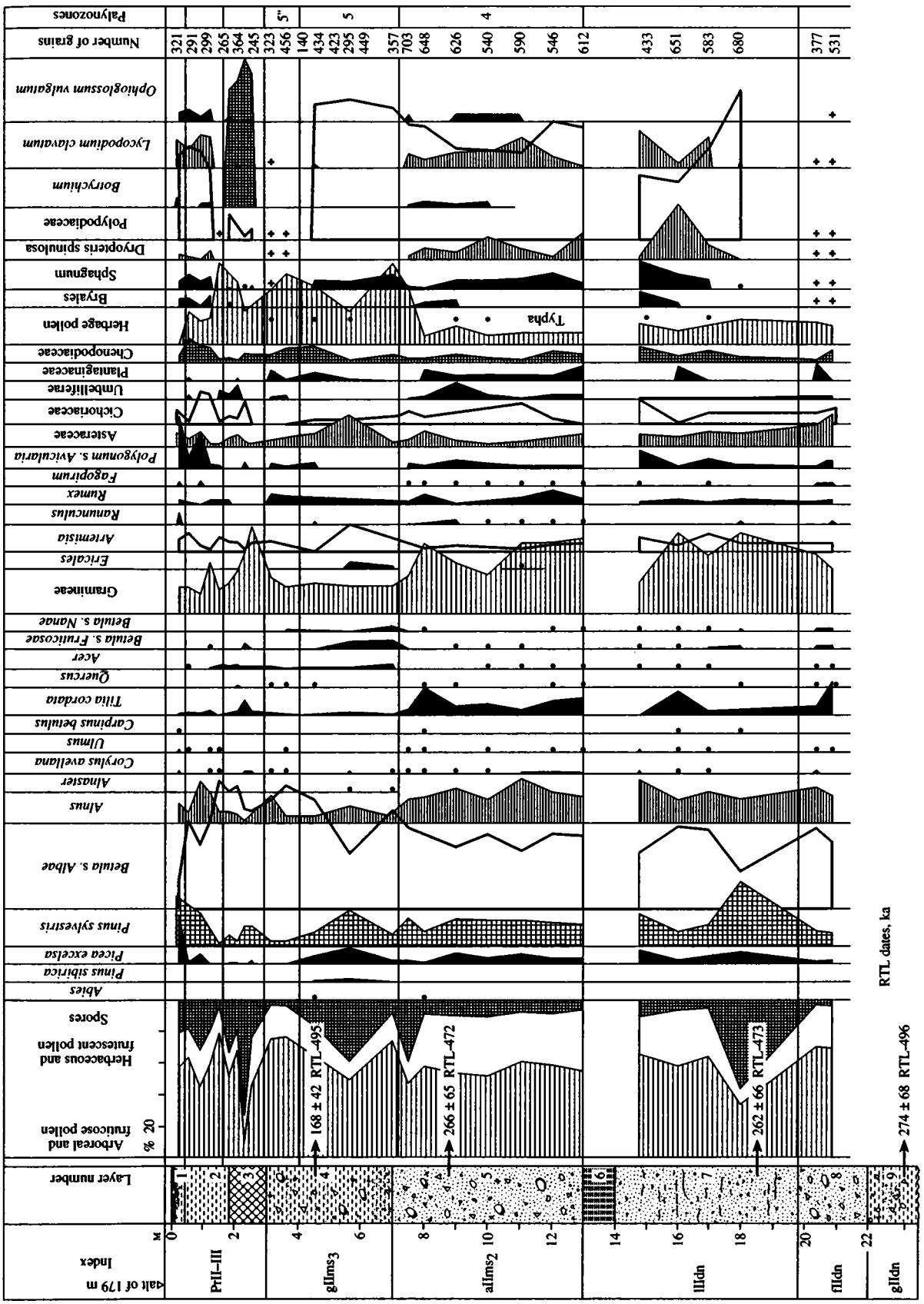

Fig. 6. Spore–pollen diagram for deposits recovered by Borehole V-8-1 (symbols as in Figs. 2 and 4).

megainterstadial rank of the middle Moskowian warming. Some researchers (Pisareva, 1965; Zelikson and Monoszon, 1981) usually considered pollen of broad-leaved taxa from interstadial sediments of the glacial horizons as redeposited. In the study area, spores, pollen, and various phytoplankton redeposited from the Paleozoic and Mesozoic beds are found only in the pre-Dnieper and, particularly, in the Dnieper sequences (*Kompleksnyi analiz...*, 1992). In opinion of E.M. Malaeva, the studied deposits of the Moskowian Horizon almost lack the redeposited microfossils except for the rare Carboniferous spores and single specimens of ancient phytoplankton.

Pollen occurring *in situ* within morainal loam beds and quite regular character of the pollen spectra elucidate specific conditions of sedimentation and burial of microfossils during the glacial epochs. On the one hand, accumulation of boulder loam by thawing of glaciers that bear morainal material suggests rather severe climatic environments, but on the other hand, fairly diverse spores and pollen buried in the glacial sediments imply presence of a much more thermophilic vegetation in the periglacial zone that is incompatible with traditional viewpoints on corresponding natural environments. One of the possible solutions of this contradiction is the supposition about the long-term character of deglaciation, when individual masses of buried ice containing morainal material thawed during a long period of time under the cover of loose deposits, on which arboreal plants already grew. The buried glaciers releasing clastic material by thawing are known in the present-day taiga landscape of northern Siberia (Kaplyanskaya and Tarnogradskii, 1976; Astachov and Isaeva, 1988; etc.), and this allows us to suggest similar environments as characteristic of the deglaciation epochs of the Pleistocene ice sheets in Europe.

The available RTL dates (Figs. 5 and 6) are relatively reliable only for the second stage of the Moskowian glaciation (moraine M1) that is about 170 ka old, and the middle Moskowian epoch can be conventionally attributed to the time interval older than 170 ka. This inference is compatible with the date of 190 ± 20 ka estimated for the middle Moskowian epoch in the Barmino section (Vladimir region) located in the extraglacial zone of the Moskowian glaciation (Zubakov, 1992).

The following conclusions can be inferred from the above discussion:

(1) The Moskowian Horizon of the middle Pleistocene area can be divided within the southwestern Moscow region into several subdivisions corresponding to the glacial and intramorainal sediments, the formation of which recorded changes in natural environments. According to geological and geomorphological data, there were four stages of the glacial sheet development: the Borovsk, Kaluga, Mozhaisk, and Krasnyi Kholm stadials.

(2) Data on the Protva River basin and the adjoining territories testify that the two first stages were divided by the longest interval of warming defined as megainterstadial. This allows us to distinguish three important periods of the Moskowian epoch: the early, middle, and late Moskowian substages. The middle Moskowian Megainterstadial also has name "Protvino" and is tentatively correlated with the Kostroma Interstadial distinguished by Pisareva (1965).

(3) Palynological characteristics of deposits of the middle Moskowian epoch show that climatic conditions during the climatic optimum were of the interstadial type, and events of that time are comparable in their paleogeographic significance with those of the middle Valdai (Monchalovo) epoch.

(4) Palynological characteristics of the morainal loams proper, in particular, the arboreal type and peculiar features of their pollen spectra, allow us to speculate about a specific (slow) character of deglaciation, when the ice masses containing morainal material could be preserved for a considerably long period under the loose deposits, which were at that time overgrown with arboreal vegetation.

Conclusions obtained after generalization of new factual data on climatostratigraphic subdivisions of the Moskowian Horizon in the Mozhaisk–Tarusa sector of the glacial sheet may be important for correlation of deposits accumulated in the second half of the middle Pleistocene in the central part of the Russian plain.

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